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GEOLOGICAL MAGAZINE

VOL. XCVI, No. 2. MARCH–APRIL, 1959

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Gravity Anomalies in the Channel Islands

By A. A. DAY

ABSTRACT

Bouguer gravity anomalies have been determined at seventy-nine stations on Alderney, Guernsey, and Jersey. On Alderney the anomalies are clearly dependent on the nature of the outcropping rock-types, and permit approximate values for the thickness of two outcropping rock masses to be obtained. On Guernsey the anomalies are not closely related to the surface geology, and suggest that the north-eastern coastal area is underlain by a body of dense rock, possibly of ultrabasic composition. The anomalies on Jersey indicate that the sedimentary rocks of western Jersey are underlain at no great depth by rock of density comparable to that of granite. In eastern Jersey the dominating feature of the anomalies is a pronounced " high " centred near Grande Charrière. It is shown that this feature is most reasonably considered to be the effect of a large buried gabbro intrusion.

I. INTRODUCTION

GRAVITY surveys of Alderney, Guernsey, and Jersey were carried out in 1953 and 1954 using the Worden gravity meter belonging to the Geological Survey. The surveys were referred to base stations in St. Anne, St. Peter Port, and St. Helier, the gravity values of which have been reported elsewhere (Day, 1955).

From a gravimetric point of view the geology of the Channel Islands is relatively complex because of the wide variety of rock-types and corresponding densities. In order that the gravity surveys should give useful information on the relationships of the various rock masses it was therefore necessary to have a density of gravity observations rather greater than that required, for example, in the south or south-west of England, where the geology is susceptible to broad generalization.

The surveys of the islands were carried out on foot. Five field stations were established on Alderney, twelve on Guernsey, and fifty-nine on Jersey. The stations on Alderney constitute a profile along the island, whereas those on Guernsey and Jersey are distributed as evenly as possible over the entirety of each island.

Bouguer anomalies were calculated in the usual way, referred to the International Gravity Formula and to the conventional base value at

the Pendulum House, Cambridge, of g = 981·2650 gal. (1 gal. = 1 cm./sec.^{2}). The anomalies were corrected for terrain effects to the nearest 0·1 milligal. The standard error of the differences between the anomalies at adjacent stations on Alderney is about ± 0·2 mgal., and on Guernsey and Jersey it is about ± 0·1 mgal.

II. Rock Densities

The densities used in the calculation of the anomalies were based on measurements of typical samples of the various rock-types of the islands, although few specimens of some types were obtainable because a thick mantle of *limon* and soil covered the solid rock. The mean values of density used in the reductions are given in Table 1 and apply to saturated specimens. A selection of individual densities was given in a previous paper (Day *et al.*, 1956, Table IV) in conjunction with the values of the velocity of compressional sound waves through the specimens.

Table 1.—Densities of Channel Islands Rock Formations (gm./c.c.)
(Values are given to the nearest 0·05 gm./c.c.)

Alderney	Cambrian grit (dry density 2·55)	2·65
	Pre-Cambrian granite-porphyry	2·65
	„ diorite	2·85
	„ gabbro	2·9
Guernsey	Granite gneiss	2·65
	Diorite gneiss	2·85
	Granodiorite } Differentiation complex	2·65
	Diorite } Differentiation complex	2·85
	Gabbro } Differentiation complex	2·9
	Granite (Cobo Bay)	2·6
Jersey	Pre-Cambrian shale and greywacke	2·75
	" Newer " granite	2·6
	" Older " granite	2·65
	Gabbro	2·9
	Rhyolite, felsite	2·65
	" Andesite "	2·8

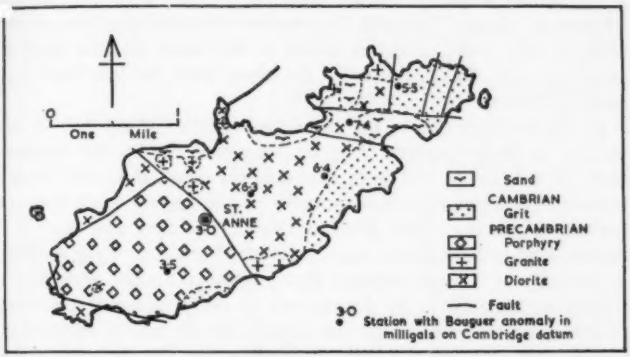
The values of density adopted are given only to the nearest 0·05 gm./c.c. because in general they depend on measurements made on specimens collected from surface outcrops. There are few active quarries in the islands and hence it was difficult to obtain undoubtedly fresh samples in all cases. Moreover, it was not possible to make any reliable determinations of formation density by the gravimetric method owing to a lack of suitable localities.

These difficulties are not unusual, but it has nevertheless been the practice of some authors to use values for formation densities with an apparent precision of about ± 0·01 gm./c.c. although they are based upon the measured densities of relatively near surface specimens. But the changes brought about by weathering and the circulation of

ground water in rocks (even including granites) which possess appreciable permeability may either lower, or raise by secondary silification, the density of these rocks. See also Cook and Murphy (1952) in this connection.

III. Interpretation of the Anomalies on Alderney

The positions of all the gravity stations, together with the Bouguer anomalies, are shown in Text-fig. 1 superimposed on a geological map of the island adapted from that published by Plymen (1922). In the higher parts of the island, where there are few outcrops, the positions of the geological boundaries are somewhat uncertain.



Text-fig. 1.—Bouguer gravity anomalies on Alderney. Datum: Pendulum House, Cambridge, g = 981·2650 gal. Geology after Plymen (1922).

The conspicuous features of the profile of anomalies are: (*a*) an increase of 3 mgal. that occurs north-eastward from St. Anne; and (*b*) the decrease of 2 mgal. between the Coastguard Station and Berry's Quarry, in the northern part of the island.

Plymen has shown that the boundary between the diorite and the granite-porphyry is a fault which probably traverses the entire width of the island and passes approximately under St. Anne, its exact course in the centre of the island being obscured. On the basis of the petrological and other characteristics of the granite-porphry mass, Plymen (1922, p. 134) and Groves (1927, p. 468) have suggested that it forms part of a sill intruding the dioritic complex.

According to the values given in Table 1, there should exist a density contrast across the boundary fault between granite-porphyry and diorite of 0·2 gm./c.c., the granite-porphyry having the lower density.

The occurrence of lower anomalies over the porphyry is therefore in qualitative agreement with the surface geology, provided that it be assumed that the boundary fault passes a little to the north-east of St. Anne, and not beneath the town in the manner suggested by Plymen The trace of the fault shown in Text-fig. 1 has been adjusted to agree with the gravitational evidence.

An approximate value may be obtained for the thickness of the sill-like block of granite-porphyry from the formula for the gravity effect of an infinite horizontal sheet of material. For this it is necessary to make the not unreasonable assumption that the porphyry is underlain by diorite. Then, with a density contrast between diorite and porphyry of 0·2 gm./c.c., and a gravity effect of 3 mgal., the thickness of the porphyry below sea level found from the formula is 360 m. Should there be a " feeder " beneath the porphyry the total observed gravity effect is due to the combined effects of the feeder and the sheet of porphyry, and the thickness of the sheet must be less than that calculated above.

At the northern end of the island the decrease of anomaly in the vicinity of Berry's Quarry may be accounted for by the presence there of sandstones and grits that are less dense than the dioritic basement on which they most probably lie and against which they are faulted. From the values given in Table 1 it is clear that there is a density contrast between grit and diorite of 0·2 gm./c.c. If the decrease of anomaly of 2 mgal. between the Coastguard Station and Berry's Quarry be assumed to be due entirely to the presence of the sheet of grit under the quarry, then the formula for the gravity effect of an infinite horizontal sheet of material suggests that the thickness of the grits is at least 240 m. This estimate is a minimum value because the total gravity effect of the grit probably attains its greatest value at points more distant from the boundary fault than the northern gravity station.

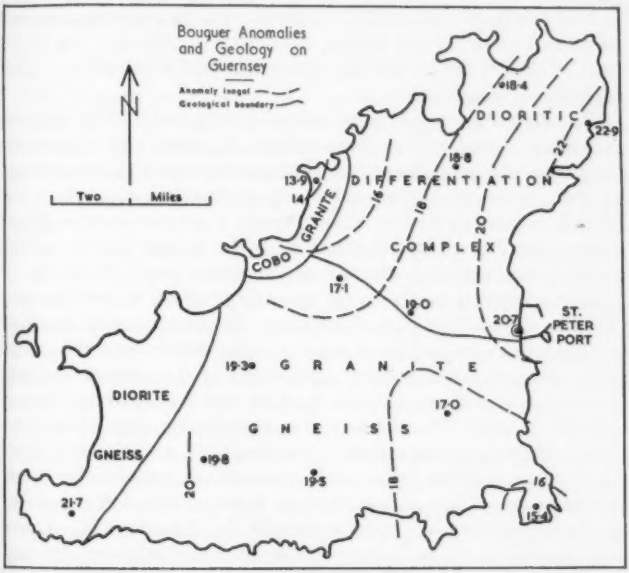
The estimated thickness of the grits beneath Berry's Quarry falls within the limits imposed by Plymen's estimate (1922, p. 128) of 1,200 ft. for the total thickness of grit exposed on the island. But the estimate is not compatible with Plymen's conclusion, based on stratigraphical evidence, that since a conglomeratic phase of the formation is exposed in the quarry the basement rock is very probably at a shallow depth beneath the quarry floor.

The alternative explanation of the reduced anomaly at Berry's Quarry—that it is due to the density of the basement beneath the grit being less than that of diorite, and possibly corresponding to that of a more acid rock-type—is improbable since the known outcrops of granite in the dioritic complex are very small in extent, and insufficient to give a gravity effect of the correct magnitude.

IV. Interpretation of the Anomalies on Guernsey

The Bouguer anomaly contours are shown in Text-fig. 2 superimposed on a geological map of the island.

The geology of Guernsey has been worked out in considerable detail by a number of investigators, of whom the latest are Plymen and Groves; the geological detail shown in Text-fig. 2 is a slightly simplified version of a map by Plymen (1932). The northern part of



TEXT-FIG. 2.—Bouguer gravity anomalies on Guernsey. Geology after Plymen (1932).

the island is occupied by the so-called Differentiation Complex, which is predominantly basic to intermediate in composition but which includes a number of small areas of acidic rocks as well as a large area of distinctive granite known as the Cobo Granite. The southern boundary of the differentiation complex is an important fault that traverses very nearly the entire island. Only the Cobo Granite appears to be unaffected by the fault, although that granite is certainly a member of, and probably the last stage of the differentiation series (Groves, 1927, p. 461). The southern half of the island

is composed of a series of ortho-gneisses together with a few small areas of para-schists. There are also some minor intrusions present, but these are not important from a gravimetric point of view.

The gravity anomalies in the northern half of the island show an increase of 9 mgal. from Grandes Rocques (on the Cobo Granite) eastwards to Bordeaux. The anomaly gradient between a point near L'Islet and Bordeaux is 1·6 mgal./km. This gradient and the regional increase of anomaly from west to east across the island suggest that the area near Bordeaux and St. Sampson is underlain by a mass of rock of relatively high density compared with the rock-types exposed on the surface (which are diorites, tonalites, and gabbros). The buried mass of rock to which the anomaly is ascribed is therefore possibly of ultrabasic composition.

The anomalies over the large area of granite gneiss in the southern part of the island are relatively uniform in value, and suggest that there are no large bodies of rock of dissimilar composition (very acid or basic intrusions, for example) lying at depths of less than a few kilometres from the surface. The anomaly is greater over the dioritic gneiss near Pleinmont, consistent with the greater density of the diorite gneiss compared with that of the granite gneiss (Table 1).

On the Jerbourg peninsula the anomaly observed is less than elsewhere on the southern part of the island. In the absence of any large development at the surface of rocks of lower density than the granite gneiss and schist, the simplest explanation of the reduced anomaly appears to be that an intrusive body of low-density acidic igneous rock lies at depth. The presence of several granite, aplite, and quartz veins in the rocks of the peninsula (Plymen, 1953, p. 268) also suggests the presence at depth of an acidic intrusion, in agreement with this interpretation. There may perhaps be a minor intrusion connecting the postulated intrusion at Jerbourg with the Cobo Granite, but the true significance of the trough in the anomalies between Jerbourg and Grandes Rocques is difficult to determine.

The simple nature of the anomaly field over Guernsey is noteworthy, and no doubt springs from the relatively close petrological relationship that exists between all the main rock-types on the island. Plymen (1951) has suggested that the gneisses and the non-foliated igneous rocks owe their origin essentially to the progressive granitization of schists of the type now exposed in the small patches at Jerbourg and Pleinmont. This theory does not appear to the author to be as consistent with the surface geology as Plymen claims. Certainly, the gravity anomalies over the southern part of the island give no indication of the existence of a large acid " intrusion " under the gneisses as might be expected on the hypothesis that the gneisses represent the higher zones of granitization. Again, if it is supposed that the whole island

is underlain by a body of rock of acidic type (comparable, say, with the Dartmoor Granite) then, by analogy with the anomalies observed over granites of that type, the anomalies on the island should be negative. This is not the case, neither is there any indication that the regional anomalies near the island have high positive values and that the anomalies on the island are much less. It is, therefore, reasonably certain that there is no large hidden acid igneous mass beneath the island. Rather it would appear that at least the northern part of the island is underlain by relatively dense material which is probably ultrabasic in composition.

V. Interpretation of the Anomalies on Jersey

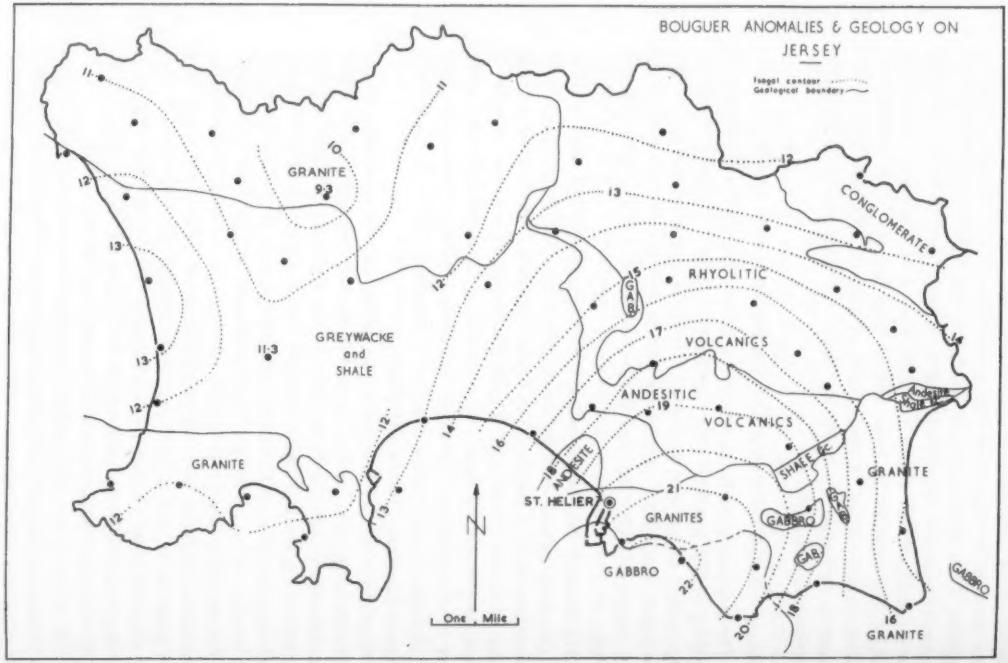
The Bouguer gravity anomalies on Jersey are shown contoured in Text-fig. 3. The geology of the island is also shown, slightly generalized. The sources of information drawn upon in the construction of the geological map were papers by Noury (1886) for the geology of the outlying reefs, Plymen (1921), Mourant (1933), and Henson (1956).

The anomaly field on the island is well determined by a large number of observations and consists essentially of two elements: an area on the western half of the island in which the anomalies have an average value of about + 11 mgal., variations above and below this value amounting to only 2 mgal.; and an area covering mainly the eastern half of the island in which the anomalies increase uniformly from + 12 mgal. on the north coast to a maximum of + 22 mgal. near Grande Charrière on the south coast.

Western Jersey.—The relationship of the gravity anomalies to the surface geology is consistent with that observed in many other areas: the anomalies over the large north-western granite complex are less than those on the adjoining stratified rocks into which it is intruded, and the anomalies on the south-western complex show a tendency to be less than those on the shale and greywacke group which it intrudes.

The anomaly gradients across the margin of the north-western granite are relatively low. Between Carrefour Selous and Trois Bois the gradient is 1·9 mgal./km., and between Lèoville House and St. Ouen's Church the gradient is 0·8 mgal./km. The total change of anomaly from north to south across the granite contact is only 2 mgal. in the vicinity of the airport and 4 mgal. near the west coast.

This evidence strongly suggests that the north-western and south-western granite masses are the surface expressions of one mass of granite, and that the Pre-Cambrian shale and greywacke group forms a relatively thin mantle over the central part of the granite mass. The contrast between the densities of the granite and the sediments (Table 1) is approximately 0·15 gm./c.c. On this basis the change of anomaly of



TEXT-FIG. 3.—Bouguer gravity anomalies on Jersey. Geology after several authors, including Noury (1886), Plymen (1921), and Mourant (1933).

2 mgal. between the north-western granite and the sediments suggests that the thickness of sediments covering the granite is about 300 m. (1,000 ft.) in the vicinity of St. Peter's Church.

The evidence of the anomalies for the existence of an underground connection between the western granites is supported by the petrological evidence. As a result of a detailed study of the two granite masses Wells and Wooldridge (1931, p. 194) concluded that they were consanguinous and are connected at no great depth below the present ground surface. It should be noted, however, that the gravity anomalies may also result from two distinct granite intrusions, possibly of different ages, with their boundaries adjoining, provided it be assumed that their bulk densities (and therefore compositions) are very similar.

Eastern Jersey.—The anomalies in this part of the island bear little relation to the distribution or structures of the surface rocks, and for their explanation require the presence of a body of high-density rock situated beneath the granite and gabbro area south-east of St. Helier.

Wells and Wooldridge (1931, p. 193) considered that the numerous outcrops of gabbro in south-eastern Jersey were the isolated remnants of a basic, sheet-like intrusion that once capped the granite complex. The large positive gravity anomaly is clearly not due to the presence of the remnants of gabbro, but it seems reasonable to suppose that it is due to the mass of gabbro in the feeder along which the magma was introduced to the higher levels.

Unfortunately the coastline cuts across the anomaly contours in such a way that it is impossible to determine whether they are approximately circular or whether they are elongated some distance southwards. In the first case the dense disturbing mass required would be approximately a circular cylinder or cone in shape, and in the second case it would be a broad vertical sheet elongated in a southerly direction. In the absence of any well-defined surface outcrop of the disturbing rock-mass, and of information concerning the complete shape of the anomaly contours, it is not possible to determine the exact shape of the disturbing mass. Its central region very probably lies below and a little to the south-west from Grande Charrière.

No conclusions may be drawn from the anomalies concerning the relationship of the south-eastern granite complex to the western complexes owing to the presence of the strong positive anomaly. It may be noted, however, that although there is a density contrast of 0·15 gm./c.c. between the granite mass and the andesitic volcanics into which it is intruded, the anomaly contours do not show any appreciable irregularity near the contact area, and consequently we may conclude that the contact-surface dips northward under the volcanics at a small inclination to the horizontal.

VI. Acknowledgments

I am grateful to the Director of the Geological Survey for granting the loan of the Survey's Worden gravity meter, and to Mr. B. C. Browne and Dr. W. Bullerwell for stimulation and encouragement. The work was carried out in the Department of Geodesy and Geophysics, Cambridge.

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Department of Geology and Geophysics,
University of Sydney.

On the Supposed Evolution of *Gryphaea* in the Lias

By A. Hallam

Abstract

A re-examination of Hettangian and Lower Sinemurian oysters has led to the rejection of Trueman's hypothesis on the evolution of *Liostrea* into *Gryphaea*. *Gryphaea* assemblages from two horizons in different parts of Britain have been subjected to bivariate and size frequency analysis. The only detectable difference is a slight size increase up the succession. No "gryphaeoid" trend is discernible in *Liostrea* up the Hettangian succession and the change to *Gryphaea* is in consequence abrupt. All the oysters of Trueman's series are referred to what are considered to be two different Linnaean species.

Introduction

In 1922 Trueman published a paper in which he claimed to have proved a continuous evolutionary series from "*Ostrea*" (now *Liostrea*) *liassica* to *Gryphaea incurva* in the basal zones of the Lias. Although more work has been done in recent years on variation in *Liostrea* and *Gryphaea* populations (Swinnerton, 1939, 1940; McLennan and Trueman, 1942) the basic picture of a progressive increase in coiling and reduction of the area of attachment of the left valve up the succession remains unchanged.

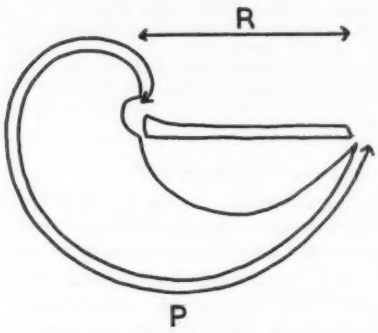
Serious doubts were aroused about Trueman's conclusions during the course of fieldwork started by the writer in 1955, so it was decided to apply modern statistical techniques to the problem. It is convenient to consider *Gryphaea* and *Liostrea* separately, in that order.

Gryphaea

Trueman expressed the degree of coiling of the left valve of *Gryphaea* as numbers of whorls. He claimed a change from about ¾ whorl in the lower *bucklandi* Zone (i.e. the *conybeari* Subzone) to 1⅛ whorls in the lower *semicostatum* Zone (*gmuendense* Subzone). Thus higher forms showed a pronounced degree of incurving (*G. incurva*) unlike the lower (*G. obliquata*). However, as Joysey (1956, 1958) has pointed out, Trueman did not take the factor of size adequately into account. The frequency distribution method that he used is unsatisfactory when adults are assessed on the basis of size, for individual gryphaeas become more coiled during growth. As Trueman also claimed an increase of size up the succession it is clear that an increase of coiling might be due entirely to such an increase. The only way of resolving this difficulty is to study the relationship between size and coiling at different horizons by means of bivariate analysis.

The amount of coiling may be assessed quantitatively by measure-

ments of the periphery (Text-fig. 1). Though this would not be satisfactory as a means of comparing *Liostrea* and *Gryphaea* it is quite suitable for the analysis in question. The periphery is measured from the distal edge of the attachment area. As the latter is nearly always less than 2 mm. in diameter this possible source of error is negligible. With *Liostrea*, however, it would be considerable. There remains the

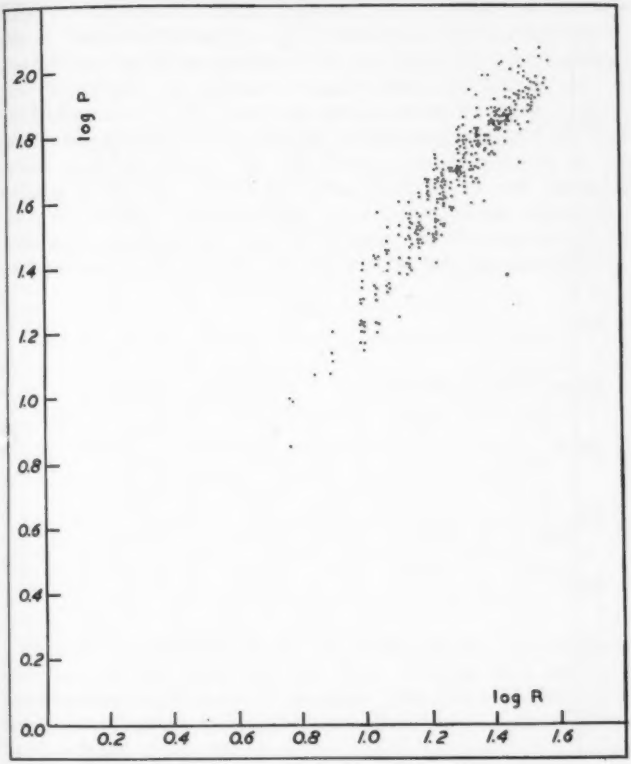


TEXT-FIG. 1.—Measurement of the periphery (P) and length of right valve (R) in *Gryphaea*.

possible objection that appreciable variations in the initial stages of coiling might be overlooked in larger specimens. If this were an important factor the smaller specimens might be expected to exhibit a greater (apparent) variability, but the scatter diagrams that have been plotted (Text-figs. 2 and 3) do not suggest this.

The length of the right valve provides a good measure of size since this corresponds closely with the length of the soft parts themselves. Right valves are not often found in association with left valves but fortunately an accurate measure can be obtained from the left valve (McLennan and Trueman, 1942, and Text-fig. 1).

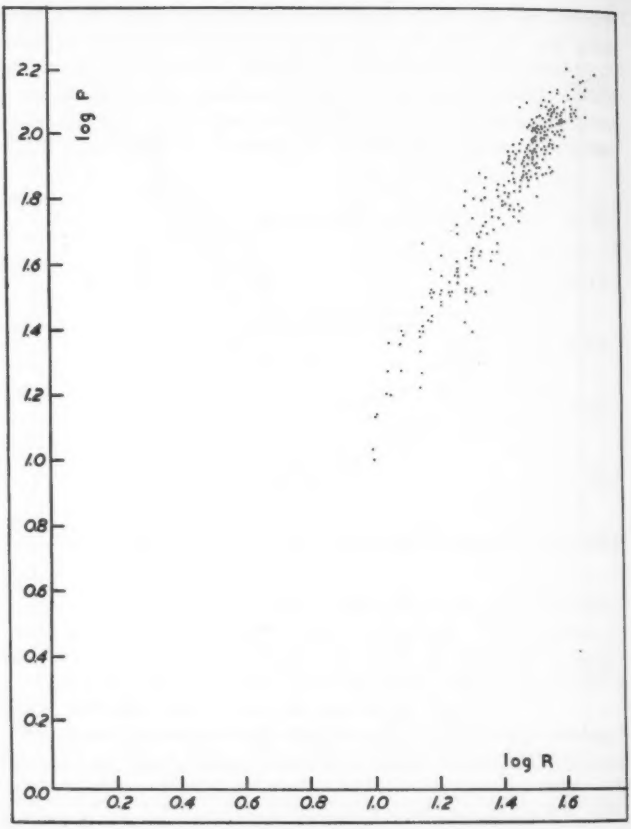
Bivariate analysis was undertaken to compare samples collected at what may be treated as two horizons. Material from the highest part of the *angulata* Zone was collected from Dorset, Somerset, Glamorgan, and Yorkshire, from the upper *bucklandi* Zone (*bucklandi* Subzone) of Yorkshire and from the lower *semicostatum* Zone (*gmuendense* Subzone) of Glamorgan and Skye. The *bucklandi* and *gmuendense* Subzones are adjacent and may conveniently be taken together. When the periphery (P) was plotted against length of right valve (R) for the two assemblages the trend of the scatter diagrams was seen to be curved, so logarithms were plotted instead, with log P as the ordinate (Text-figs. 2 and 3). This time the scatter diagrams approxi-



TEXT-FIG. 2.—Scatter diagram to show the relationship between the logarithms of the periphery and length of right valve, measured in mm., in *Gryphaea* sample from the *angulata* Subzone.

mate closely to straight lines of the form $\log P = a \log R + b$, where " a " is the slope and " b " the intercept on the ordinate. The equation may be rewritten $P = bR^a$ to express the direct relationship between P and R.

Trueman's claims can now be subjected to a simple test. The slope " a " gives information on the tightness of coiling. If Trueman is to be proved right " a " should be significantly greater for the higher horizon. " a " and " b " were calculated using the Reduced Major Axis method of Kermack and Haldane (1950). The relevant statistics are given in Table 1, p. 103.



TEXT-FIG. 3.—Scatter diagram cf. Text-fig. 2 for sample from the *bucklandi* and *gmuendense* Subzones.

It was found that " a " and " b " do not differ significantly for the two assemblages ($P > 0{\cdot}05$). Therefore that part of Trueman's hypothesis dealing with the evolution of *Gryphaea* itself is not borne out by statistical analysis. For all the differences there are between them both samples could have come from the same population.

It was noted above that a combination of ontogenetic increase of coiling and size increase up the succession might suffice to account for an apparent evolutionary change. It is therefore desirable to examine the evidence for such a size increase. The reliability of

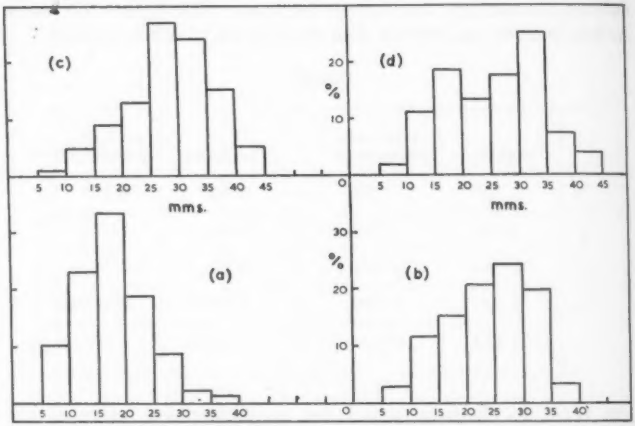
subjecting the samples to size frequency analysis must first be considered. Joysey (1956) lists six ways of bias influencing fossil collections. For the purposes of comparison of the two assemblages under discussion, which need not represent original living populations and for which significant current sorting can be discounted, the only one which might raise problems is (v), that the palaeontologist is more likely to collect large specimens. An unconscious bias on the part of the collector is not easy to avoid. However, in the present case, involving the same collector working under more or less similar conditions, differential errors are not likely to be great. Nevertheless the results are bound to be less reliable than those of the bivariate analysis.

TABLE 1

	angulata	*bucklandi-gmuendense*		*angulata*	*bucklandi-gmuendense*
N	338	276	r	0·965	0·959
$\bar{y}$	1·617	1·842	a	1·61	1·57
$\bar{x}$	1·315	1·422	b	− 0·50	− 0·39
s_y	0·224	0·236	σa	0·02298	0·02480
s_x	0·139	0·150	σb	0·04440	0·05475

In order to eliminate errors due to geographical variation the samples were split into four, two from Yorkshire and two from south-western England and Glamorgan. Size frequency histograms are given in Text-fig. 4. It is clear from these that only the Yorkshire *bucklandi* and south-western *angulata* have reasonably comparable shapes, the other two containing high proportions of juveniles, giving a strongly left-skewed histogram in one case and a bimodal histogram in the other. For the Yorkshire *bucklandi* sample $\bar{x} = 28{\cdot}9 \pm 9{\cdot}92$ and for the south-western *angulata* $24{\cdot}4 \pm 9{\cdot}75$ mm. The difference in means is not significant ($P > 0{\cdot}05$). It would be wrong, however, to deny any size increase up the succession in this particular case. If instead of means, maximum sizes are compared, some interesting results are obtained. The maximum lengths of the right valve in the two *angulata* Zone samples of Yorkshire and the South-West are 37 and 38 mm. and for their higher horizon equivalents 45 and 46 mm. respectively. Thus geographical variations seem negligible. It is interesting to note that the ratio of means, $28{\cdot}9/24{\cdot}4 = 1{\cdot}184$, is similar to the ratio of maximum sizes, $4{\cdot}6/3{\cdot}8 = 1{\cdot}211$. Given a fair number of adult specimens, then, determination of maximum size

can prove most useful. Size indeed appears constant over a wide area. For example, specimens collected by the writer in Skye from the *semicostatum* Zone extend up to 50 mm. while the bulk of the specimens collected from a similar horizon near Loch Aline by McLennan and Trueman (1942) range between 40 and 50 mm. Gryphaeas collected from the well-known locality of Fretherne, Gloucestershire (*semicostatum* Zone) have been examined in the Sedgwick and Bristol University Museums. Again, they range mostly between 40 and



TEXT-FIG. 4.—Size-frequency histograms of four *Gryphea* samples from (*a*) *angulata* Subzone, Yorkshire; (*b*) *angulata* Subzone, Dorset, Somerset, and Glamorgan; (*c*) *bucklandi* Subzone, Yorkshire, and (*d*) *gmuendense* Subzone, Glamorgan.

50 mm. Although the Württemberg Lias has been examined more cursorily than the British enough data has been collected to suggest comparable size changes from the *angulata* to the *semicostatum* Zones.

Because of this small size increase up the succession there tends to be an increased proportion of incurved forms at higher horizons. This, however, is hardly sufficient to create the impression of an evolutionary series if the samples considered are taken at random. For example, Trueman considered that *G. incurva* did not occur below the *gmuendense* Subzone, though accepting that quite curved forms occurred somewhat rarely in the *angulata* Zone and not uncommonly in the lower part of the *bucklandi* Zone. In fact most of the adults from the top *angulata* Zone are, to the present writer, indistinguishable from the typical *G. incurva*.

Liostrea

Trueman claimed a gradual change in the oysters throughout the Hettangian, from " *Ostrea liassica* " at the base of the *planorbis* Zone, with the whole surface of the left valve attached, to " *Gryphaea dumortieri* " in the *angulata* Zone, with not usually more than one-sixth of the length of the valve attached.

There are three reasons why statistical methods are hard to apply satisfactorily in this case.

(i) It is frequently difficult to obtain an accurate measure of the character in question, the area of attachment, especially when the oysters have been moulded on ribbed ammonites, as is often the case.

(ii) There is a bias in favour of collecting specimens with comparatively small areas of attachment.

(iii) Many of the shells break easily so that the collection of a large number of whole specimens is a much more trying task than with the comparatively robust gryphaeas.

It is possible, though, to make some remarks based upon extensive collecting and field observations.

A number of species of Hettangian oysters of the genus *Liostrea* have been proposed on the basis of shell shape. This character, however, is notoriously variable within the same species of living oysters, depending principally on the nature of the substratum (Orton, 1937; Gunter, 1950). As zoologists classify oysters on the basis of soft parts palaeontologists are posed with serious problems. In these circumstances a highly conservative approach is the only tenable one in the classification of Liassic oysters, at least until an elaborate taxonomic analysis involving statistics has been carried out.

All the Hettangian oysters can be referred satisfactorily to one Linnaean species, *Liostrea irregularis* (Münster). The subcentral muscle scar indicates that this species has affinities with *Ostrea* rather than *Crassostrea* among living oysters (Gunter, 1950). Swinnerton (1939, fig. 2) gives a good idea of the wide variation within *L. irregularis*. In the majority of cases the attachment area varies from about three-quarters to one-fifth of the total area of the left valve. Those forms with an attachment area of less than one-fifth appear more elongate and convex than the rest. This " gryphaeoid " trend is readily intelligible if comparison is made with living oysters. If the attachment area is small so that the oyster has to grow in soft mud it tends to be cup-shaped (Dall, 1898; McLennan and Trueman, 1942), this being a simple adaptation to ensure that the siphons are kept well above the surface of the sediment. Oysters crowded together on a mud bottom tend to be elongate (Gunter, 1938). Both the strongly curved " *Gryphaea* " *dumortieri* of the *angulata* Zone and the smaller, flatter

Liostrea hisingeri Nilsson (= *liassica* Strickland) can be considered as mere phenotypic variations of *L. irregularis*.

Careful observations all over Britain and in southern Germany have failed to reveal the slightest suggestion of a trend of the sort Trueman claimed. According to him the *Ostrea* Beds at the base of the *planorbis* Zone contain forms (*O. liassica*, p. 265) usually attached by the whole or nearly the whole surface of the left valve. In actual fact the attachment area is nearly always less than one-sixth and usually much less. This is related to the paucity of shell fragments of moderate size to form an adequate hard substratum. Higher up in the succession, in the *johnstoni* Subzone and *angulata* Zone, there is a much higher proportion of typical *L. irregularis*, a fact essentially related to the greater abundance of large shells in the rocks. Thus if it were possible to undertake a satisfactory statistical analysis the results might well suggest Trueman's story in reverse !

Again, it is difficult to avoid the conclusion that only highly selective collecting could lead one to infer a gryphaeoid trend up the succession. Unfortunately Trueman's original specimens are not available for re-examination.

There is evidence, though, of a size increase up the succession as in the case of *Gryphaea*. This must be considered in relation to changes in other elements of the fauna and will be dealt with elsewhere.

Distinction between *Liostrea* and *Gryphaea*

In the great majority of specimens morphological distinction is an easy matter. In *Gryphaea* the attachment area is usually less than 2 mm. in diameter. The *obliquata* stage is normally attained when the right valve is only 15 mm. long, and the left valve of the adult shell is incurved over the umbo of the right valve. Some difficulty may arise in two cases.

(i) *Liostrea irregularis* var. *dumortieri*, with a minute attachment area, might be confused with the *obliquata* stage of *Gryphaea*. The latter, however, has a pronounced umbonal thickening not found in *Liostrea*.

(ii) Exceptionally specimens with a large attachment area are found in association with normal *Gryphaea* assemblages. There is no reason to assume that these are not true Gryphaeas. It must be admitted that in some cases these cannot be distinguished from certain varieties of *Liostrea irregularis*, a point made also by Joysey (1958).

The correct name for the Lower Sinemurian Gryphaeas is *G. arcuata* Lamarck. This has priority over Sowerby's *G. incurva*. The names *obliquata* and *incurva* might be retained to label stages of development.

Both *Liostrea irregularis* and *Gryphaea arcuata* have a certain stratigraphical value. The former, very common in the Hettangian,

is sharply replaced by the latter just below the top of the *angulata* Zone. The change can be studied conveniently in Dorset, west of Lyme Regis. *Gryphaea arcuata* first enters the succession in bed 13 (numbering of Lang, 1924) and in bed 15 is found in great abundance associated with *Schlotheimia lymensis* (Spath). In the overlying Sinemurian *G. arcuata* is one of the commonest fossils whereas *L. irregularis* is found only spasmodically, e.g. in association with fossil driftwood. The virtual replacement of one species by the other is almost as striking as the parallel change in the ammonites, with *Coroniceras* succeeding *Schlotheimia*. The change, moreover, appears to take place at about the same horizon throughout North-West Europe. The *Ostrea* Beds at the base of the Lias form a distinctive facies in most of the English outcrops but the typical oyster, *L. irregularis* var. *hisingeri*, must be used with caution as a stratigraphical marker for it is the common form in the *angulata* Zone of the Skye area.

To conclude, *Liostrea irregularis* and *Gryphaea arcuata* are distinctive Linnaean species, i.e. they differed genotypically. While the presumption must remain that the Jurassic *Gryphaea* evolved in some way from an *Ostrea*-like ancestor a convincing evolutionary series has yet to be demonstrated.

Acknowledgments

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The specimens on which this work is based have been presented to the Sedgwick Museum.

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GRANT INSTITUTE OF GEOLOGY,
EDINBURGH.

The Mechanics of Oblique Slip Faulting

By M. H. P. Bott

Summary

The various mechanisms which could cause oblique slip faulting are briefly reviewed. It is thought that such faulting may frequently arise from the existence of preferred planes of fracture within the rocks. The dynamics of this mechanism is studied in some detail and an expression is obtained for the first direction of slip within the plane under the influence of a general stress system. For a stress system of given orientation it is found that the initial slip may occur in any possible direction within the plane, the direction depending on the relative values of the three principle pressures. The theory suggests that when a pre-existing fault is subjected to a reorientated stress system (typical or rotated) the movement after fracture will usually be oblique. In conclusion, the general implications of the theory are discussed.

I. Introduction

In a recent article in the *Geological Magazine*, Alwyn Williams (1958) has mentioned two possible types of mechanism which could give rise to oblique slip faulting. The first of these depends on a rotation of the typical crustal stress system, with horizontally and vertically orientated principal pressures, to give an obliquely orientated stress system; consequently the direction of movement within the plane of fracture would also be oblique. Williams has investigated in some detail the direction of slip for fracture under the influence of an oblique stress system and has applied the results to account for the oblique slip faulting within the Ordovician, Silurian, and Old Red Sandstone rocks near Girvan. The second type of mechanism depends on inhomogeneity of either strength or elastic properties within the rocks.

The dynamics of these two mechanisms are discussed in this paper. In particular the theory of inhomogeneity of strength is developed as it is thought that this mechanism may be of widespread importance. A notable feature of the theory is the controlling influence of the relative value of the intermediate principal pressure on the direction of the initial slip within the fault plane.

II. The Mechanics of Rotation of the Stress System

A typical crustal stress system, of the type envisaged by Anderson (1951), may be defined by horizontally and vertically orientated principal pressures. "Rotation" of a typical stress system can occur through combination with an obliquely orientated supplementary stress system (Hafner, 1951).

Perhaps the commonest source of obliquely orientated stress systems

within the earth's crust is provided by major inhomogeneities of density within or below the crust. It may be shown, for instance, that considerable stress differences are associated with post-tectonic granites of characteristically lower density than the intruded country rocks (Bott, 1958) ; their effect is to modify the existing typical crustal stress system in both magnitude and direction. The approximate amount of rotation of a given typical stress system may be roughly obtained by calculating the stress differences caused by the anomalous " load " on the crust (Jeffreys, 1952, Chapter 6) and combining with the typical system. The order of magnitude of the load can be obtained from the gravity anomalies. In the investigation mentioned above it was found that a fairly large post-tectonic granite, giving a gravity anomaly of about − 40 mgal., could rotate a crustal stress system large enough to initiate faulting to an extent of about 10 degrees maximum. More extensive rotation would be expected within orogenic belts or perhaps as a result of sub-crustal forces of obscure origin.

It would appear, then, that a large rotation of the typical stress system as a result of interaction with a supplementary stress system could be caused by mass distributions giving rise to major uncompensated gravity anomalies or by sub-crustal forces of similar magnitude. These conditions are likely to be found during orogenic rather than epeirogenic movements. The main conclusion is that normally only a limited amount of rotation would occur in regions suffering epeirogenic deformation.

III. Types of Rock Inhomogensity

Two fundamentally different types of inhomogeneity are liable to influence the structural behaviour of rocks. These are discussed separately below.

(*a*) *Inhomogeneity of Elastic Properties*

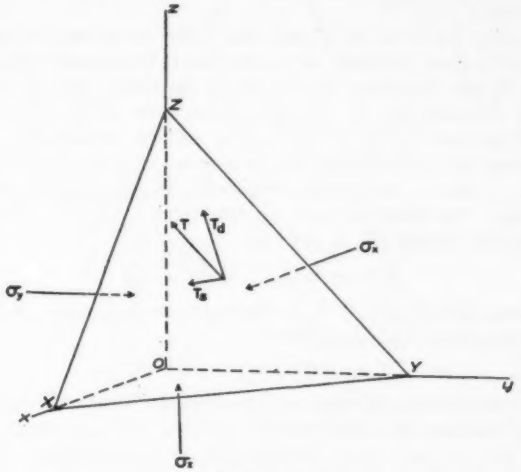
Elastic properties usually vary from one lithology to another, as may be shown by a combined study of elastic wave velocities and densities. A quantitative investigation of the influence of elastic inhomogeneity in rocks on a stress system has apparently not been made : it would appear, however, that some increase of stress would be expected within the less compressible and more rigid rock and that some rotation may occur, particularly near lithological boundaries.

Another possible cause of elastic inhomogeneity, possibly of lesser importance, arises from preferred orientation of constituent minerals within an otherwise homogeneous rock.

(*b*) *Inhomogeneity of Strength*

Inhomogeneity of strength depends on an inherent weakness of the

rocks within certain preferred directions. Such a preferred direction of weakness is likely to result from variation in lithology as different lithologies are likely to have different strengths as well as different elastic properties. The plane of weakness should usually lie parallel to the bedding planes. But preferred planes of weakness may also be found within otherwise homogeneous (or nearly homogeneous) masses of rock in the form of earlier faults, joints, or cleavage. The existence of these latter types should not noticeably affect the elastic properties and therefore the stress system would remain unaffected while elastic



TEXT-FIG. 1

conditions held. These preferred planes would remain "unnoticed" until the shearing stress within them should exceed the strength. If homogeneous fracture had not by then taken place within the rock, fracture would occur within the preferred plane in which the strength was first exceeded, and the direction of the initial slip would be defined by the direction of the greatest shearing stress within the plane.

The dynamics of simple inhomogeneity of strength within otherwise homogeneous rocks, as a mechanism of oblique slip faulting, is investigated in the following paragraph.

IV. The Dynamics of Fracture within a Preferred Plane

It is evident that the direction of movement immediately after fracture should coincide with the direction of maximum shearing stress within the fracture plane immediately before. In this paragraph

we derive an expression for the direction of maximum shearing stress within a given plane subjected to a general stress system, which will thus also give the expected direction of faulting.

The co-ordinate axes (x, y, x) are taken parallel to the corresponding principal pressures (σ_x, σ_y, σ_z). We require to find the direction of maximum shearing stress within a general plane having direction cosines (l, m, n), identically connected by the relation :—

$$l^2 + m^2 + n^2 = 1 \quad . \quad . \quad . \quad . \quad (1)$$

In Text-fig. 1 the plane is shown intercepting the co-ordinate axes at X, Y, and Z.

We take ΔXYZ to be of unit area. The component of pressure acting in a given direction across the face is then equal to the force acting in that direction. It can readily be shown that $\Delta OYZ = l$, $\Delta OZX = m$, $\Delta OXY = n$. The normal forces on ΔOYZ, ΔOZX, and ΔOXY are then found to be $l\sigma_x$, $m\sigma_y$, and $n\sigma_z$ respectively. Since the system is in equilibrium the components of force on ΔCYZ are $(-l\sigma_x, -m\sigma_y, -n\sigma_z)$, acting parallel to the corresponding co-ordinate axes. Therefore the total resultant force exerted on ΔXYZ by the material outside, F_t, is given by :—

$$F_t = -\sqrt{l^2\sigma_x^2 + m^2\sigma_y^2 + n^2\sigma_z^2}$$

the normal force on ΔXYZ, F_n, is obtained by resolving each component along the normal direction as follows :—

$$F_n = -(l^2\sigma_x + m^2\sigma_y + n^2\sigma_z)$$

The maximum shearing stress acting within the surface ΔXYZ, T_{max}, may be calculated from the relation $F^2_t = F^2_n + T^2_{max}$. Hence

$$T^2_{max} = l^2\sigma_x^2 + m^2\sigma_y^2 + n^2\sigma_z^2 - (l^2\sigma_x + m^2\sigma_y + n^2\sigma_z)^2 \quad . \quad (2)$$

Combining (1) and (2),

$$T^2_{max} = m^2(\sigma_y - \sigma_x)^2 + n^2(\sigma_z - \sigma_x)^2 - \{m^2(\sigma_y - \sigma_x) + n^2(\sigma_z - \sigma_x)\}^2 \quad . \quad . \quad . \quad (3)$$

For convenience we now consider the z axis to be vertical. We are aiming to calculate the resolved parts of T_{max}, the force per unit area exerted by the material above the plane on the material below the plane, up the line of greatest slope (T_{dip}) and along the horizontal direction YX (T_{str}). These components are connected by the following relation :—

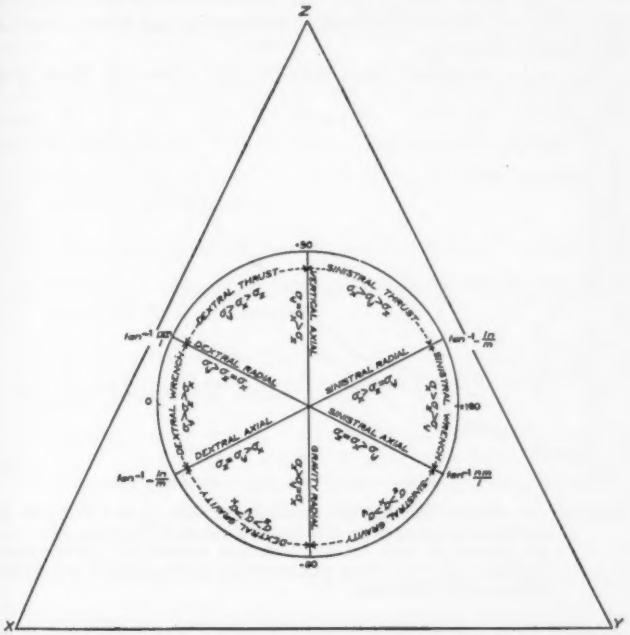
$$T^2_{max} = T^2_{str} + T^2_{dip} \quad . \quad . \quad . \quad (4)$$

Resolving the components of force along the direction YX, it may be shown that :—

$$T_{str} = \frac{lm\ (\sigma_y - \sigma_x)}{\sqrt{l^2 + m^2}} \quad . \quad . \quad . \quad . \quad (5)$$

Combining (3), (4), and (5) we obtain an expression for T_{dip}, which simplifies, on use of (1), to give the following relation :—

$$T_{dip} = \frac{n\{m^2(\sigma_y - \sigma_x) - (1 - n^2)(\sigma_z - \sigma_x)\}}{\sqrt{1 - n^2}} \quad . \quad (6)$$



TEXT-FIG. 2.—Diagram showing the direction of maximum shearing stress within the plane of preferred fracture under all possible stress regimes. The preferred plane is taken as striking at 45° to the two horizontal principal pressures and as dipping at 60°.

The pitch of the maximum shearing stress (the angle within the plane between the strike direction and the direction of maximum shearing stress), θ, is given by the relation :—

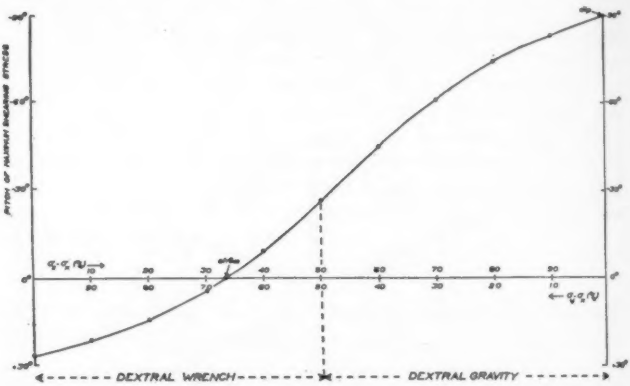
$$\tan \theta = T_{dip}/T_{str}$$

Substituting for T_{dip} and T_{str} from (5) and (6) :—

$$\tan \theta = \frac{n}{lm}\left\{m^2 - (1 - n^2)\frac{\sigma_z - \sigma_x}{\sigma_y - \sigma_x}\right\} \quad . \quad . \quad . \quad (7)$$

Under the sign convention adopted for T_{dip} and T_{str} an angle of zero corresponds to a horizontal force in the direction YX exerted by the material above the face on the material below the face. A purely dextral strike slip movement would occur after fracture in these circumstances. Positive angles (0° to 180°) involve a thrust component of dip slip after fracture, and are measured clockwise. Negative angles (0° to − 180°) are measured anticlockwise and involve a normal component of dip slip.

The main conclusion which may be drawn from the formula of



TEXT-FIG. 3.—Graph showing the variation of pitch of the direction of maximum shearing stress as the relative values of the stress differences are allowed to vary, within the dextral wrench and gravity stress regimes. The two stress differences are expressed as a percentage proportion of their sum.

equation (7) is that by unrestricted variation of the relative values of the three principal pressures the direction of maximum shearing stress within the plane may lie in every possible direction. Thus oblique slip faulting may occur in any possible direction for any given orientation of the three principal pressures. The only exceptions occur when the plane of fracture lies parallel to one of the principal pressures.

It is convenient to discuss the general behaviour of the direction of maximum shearing stress (and thus also of initial slip) in terms of tectonic stress regimes (Harland and Bayly, 1958). The concept of tectonic regimes has been mainly concerned with orogenic deformation and the movement needs to be orientated with reference to the orogenic belt so giving the primary and secondary regimes of Harland and Bayly. For faulting by itself, however, the fault plane provides the only

convenient reference, and the primary and secondary regimes are better replaced by dextral and sinistral regimes (related to the plane of fracture). As an example a plane dipping at 60° and striking at 45° to each of the horizontal principal pressures is taken for study. The disposition of the direction of maximum shearing stress for all possible stress regimes for this model is illustrated diagrammatically in Text-fig. 2, which shows a plan diagram of the surface *XYZ*. The variation of pitch of the direction, under dextral wrench and dextral gravity regimes, is illustrated graphically in Text-fig. 3.

The general behaviour of the direction of maximum shearing stress under different stress regimes adapted from Harland and Bayly is summarized as follows, Text-fig. 2 providing a specific example :—

(i) *Gravity radial regimes*, $\sigma_z > \sigma_y = \sigma_x$. $\theta = -90°$, resulting in dip slip normal faulting.

(ii) *Dextral gravity regime*, $\sigma_z > \sigma_y > \sigma_x$. $-90° < \theta < -\tan^{-1}\frac{ln}{m}$, resulting in oblique slip faulting with normal dip slip and small dextral strike slip components.

(iii) *Dextral axial regime*, $\sigma_z = \sigma_y > \sigma_x$. $\theta = -\tan^{-1}\frac{ln}{m}$, with a movement pattern similar to (ii).

(iv) *Dextral wrench regime*, $\sigma_y > \sigma_z > \sigma_x$. When $(\sigma_z - \sigma_x)/(\sigma_y - \sigma_x) = m^2/(1 - n^2)$, $\theta = 0°$, resulting in horizontal dextral movement. If the ratio of stress differences is greater than $m^2/(1 - n^2)$ there will be a normal component of dip slip, and if less there will be a thrust component of dip slip.

(v) *Dextral radial regime*, $\sigma_y > \sigma_z = \sigma_x$. $\theta = \tan^{-1}\frac{mn}{l}$, resulting in oblique slip faulting with thrust dip slip and dextral strike slip components.

(vi) *Dextral thrust regime*, $\sigma_y > \sigma_x > \sigma_z$. $\tan^{-1}\frac{mn}{l} < \theta < 90°$, resulting in thrust dip slip with a component of dextral strike slip.

(vii) *Vertical axial regime*, $\sigma_y = \sigma_x > \sigma_z$. $\theta = 90°$, resulting in simple dip slip thrust faulting.

In the summary above $\sigma_y \geqslant \sigma_x$, so that the component of strike slip is always in the dextral sense. If the numerical values of σ_x and σ_y are interchanged a corresponding pattern of movement with sinistral components of strike slip are obtained, giving a further five regimes. Thus a total of twelve regimes correspond to regimes 2 to 13 inclusive of Harland and Bayly.

V. Discussion

It has been shown in the previous paragraph that the maximum shearing stress within a preferred plane of fracture—and thus the initial slip direction after fracture—may lie in every possible direction for a variable stress system of given orientation, provided the plane is not parallel to any of the principal pressures. This conclusion is valid whatever the orientation of the stress system. Thus it is unnecessary to postulate rotation of the typical stress system to account for oblique slip originating in this way, although such slip may equally well occur in any direction under the influence of an oblique stress system.

There are two conspicuous contrasts with the theory of faulting from fracture in a homogeneous medium, as developed by Anderson (1951). In the theory of homogeneous fracture the relative value of the intermediate principal pressure has no bearing on the direction or incidence of fracture: the direction is, however, controlled by the internal coefficient of friction of the rock. In the theory of inhomogeneous strength as developed above, once the directions of the maximum and minimum principal pressures are defined, the direction of the initial slip depends only on the relative value of the intermediate principal pressure: and the only influence of the internal coefficient of friction would be to delay fracture—a delayed fracture may involve an alteration of the relative value of the intermediate principal pressure during the build-up of stresses, and thus a slightly modified initial slip direction.

Perhaps the commonest type of oblique slip of this type is likely to be found in renewed movement along pre-existing fault planes under the influence of a reorientated stress system. We should expect the first movement to conform to one of the classical classes of fault, while subsequent movements would in general show oblique slip. Examples of such renewed movement are common, frequently being revealed by slickensiding.

The theory also gives us the chance of determining the orientation of a stress system causing a given system of movement together with the relative value of the intermediate principal pressure, provided that the direction of oblique slip can be obtained for three or more different preferred directions of fracture.

Perhaps the theory may be most simply applied to subsequent movements on vertical joint surfaces. Under a typical stress system only horizontal slickensiding should be found according to the predictions of the theory, although oblique slip should be characteristic of a rotated system. This criterion should enable us to determine whether or not the stress system has been rotated when slip occurred.

It is possible that faulting commonly occurs under the influence

of a mechanism intermediate between homogeneous and inhomogeneous fracture. A set of circumstances favourable to such a mechanism would occur when a preferred direction of strike for the fracture is favoured by a basement fracture pattern without a definite control of the hade of the fracture plane. When the horizontal principal pressures lie obliquely to the strike, oblique slip faulting would be expected. The hade of the fault plane would be influenced by the internal coefficient of friction as well as by the orientation and relative magnitude of the principal stresses. Primary oblique slip faults, either affecting the basement or the overlying sediments, could originate in this way. Perhaps basement control of overlying fracture patterns is of greater importance in the development of subsequent faulting than has generally been realized.

Apparent oblique slip faulting may sometimes occur as a resulting of tilting, but the writer follows Williams in rejecting this as the most important mechanism, because frequently a multitude of different tiltings would be required to account for the variety of slickenside directions in a region. Oblique slip faulting may also take place by fracture within a homogeneous mass of rocks under the influence of a rotated system. Under usual circumstances, however, it is thought that only a limited amount of rotation can occur, and that preferred fracture planes and directions provide the usual mechanism. In conclusion, it seems possible that the oblique slip faults of the Girvan district may also be explained as a single period of inhomogeneous fracture under the influence of an approximately north–south horizontal tension, with possible variation of the intensity of the intermediate pressure.

I should like to express my gratitude to Dr. R. A. Smith for checking and suggesting improvements to the mathematical side of the paper.

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DEPARTMENT OF GEOLOGY,
SCIENCE LABORATORIES,
DURHAM.

Geological Interpretation of Airborne Magnetometer Observations from Nelson to Waikato River, New Zealand

By H. W. Wellman

Abstract

A map is presented showing a series of airborne magnetometer profiles from the north end of the South Island of New Zealand to the mouth of Waikato River. The two southern profiles are related to the exposed Upper Palaeozoic igneous rocks which are considered to extend north across Cook Strait and along the west coast of the North Island to cause the anomalies in the northern profiles. The North Island profiles are considered to reflect the Kawhia Syncline and a major anticline to the east. The eastward displacement of the magnetic low relative to the synclinal axis at the surface is considered due to the eastward dip of the axial plane of the syncline.

I. Introduction

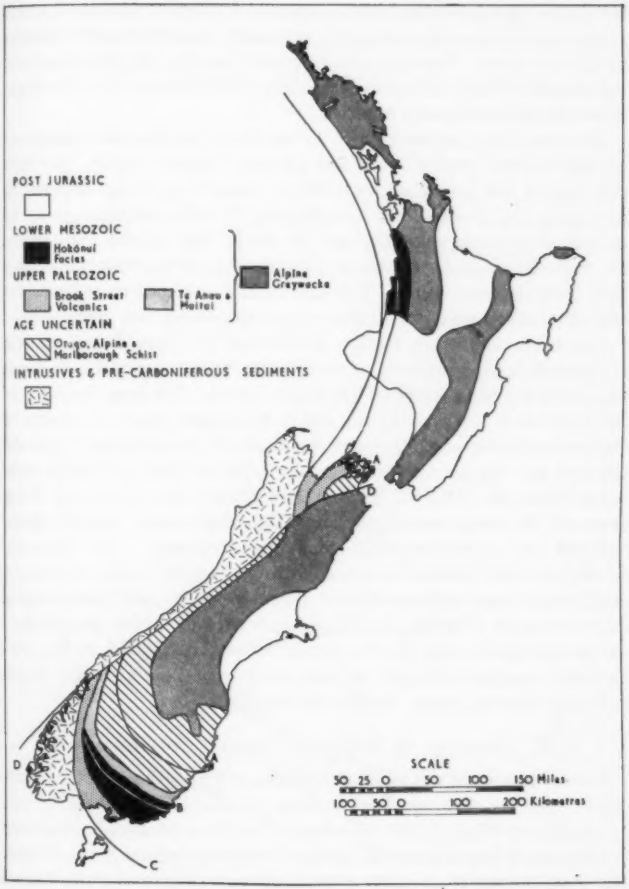
EXTREMELY thick basic lavas and associated ultra-basic rocks are interbedded with Upper Palaeozoic sediments in the South Island of New Zealand. The igneous rocks dip steeply, extend continuously for many miles, and cause large magnetic disturbances that have been known for a long time (Farr, 1916). The disturbances have recently been traced across Cook Strait and along the west coast of the North Island by airborne magnetometer observations (Gerard and Lawrie, 1956).

The airborne magnetometer is the ideal instrument for such observations. It is rapid and independent of terrain, observations being made as easily over mountains and over the sea as over level plains. A further advantage is that the magnetic pattern is viewed from a height of several thousand feet, and the magnetic disturbances are " seen " in better perspective than when viewed from the ground.

The Upper Palaeozoic volcanics are part of an essentially conformable upper Palaeozoic and lower Mesozoic sequence that accumulated within the New Zealand Geosyncline (Mutch, 1957), a major feature that extended north from New Zealand probably to New Caledonia. The western margin lay near the present southern and western coasts of the two main islands of New Zealand. The position of the eastern margin is unknown. The sediments were deformed during the growth of the geosyncline and more intensely later, the degree of deformation increasing to the east away from the margin. The magnetic rocks dealt with in this account lay near the western margin (Text-fig. 1), and although they dip steeply, the structure is reasonably simple and the stratigraphy fairly well known.

II. Magnetic Observations

Aeromagnetic surveys were made from 1949 to 1952. The results were analysed by Gerard and Lawrie and published in 1955. The



Text-fig. 1.—Locality map showing main elements of pre-Cretaceous structure of New Zealand.

A. Schist Axis.
B. Axis of major syncline.
C. Hypothetical margin of New Zealand Geosyncline.
D. Alpine fault.

main thermal area of the North Island was closely observed and widely spaced lines flown over the central and southern parts of the North Island and Cook Strait. A single line was flown along the east coast of the South Island as far south as Christchurch, and two lines flown between Christchurch and Mernoo Bank. In order to obtain absolute values the flights were made over previously observed ground stations and tied to them. The eight widely spaced lines that provide structural information on the rocks between Nelson and the mouth of Waikato River are discussed in this account.

The basic data were presented as total force profiles and generalized on two contour maps (Gerard and Lawrie : Maps 5 and 6). The first map shows the smoothed total values, smoothing being effected by averaging the profiles over overlapping 20 mile intervals so as to provide an average value for each 10 miles. The second map shows the regional magnetic anomalies obtained by subtracting the earth's field from the total values, The earth's field, assumed to be linear for latitude and longitude, was based on smooth profiles over greywacke.

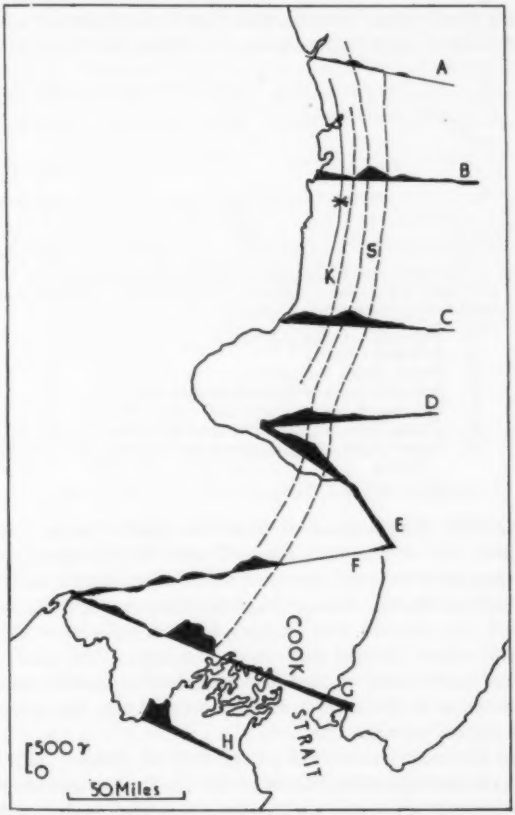
The most important feature discovered by Gerard and Lawrie is the well-defined positive anomaly from Nelson across Cook Strait and along the West Coast of the North Island. The large anomaly of 200 gammas is almost certainly due to the Upper Palaeozoic rocks of Nelson extending north into the North Island, in agreement with the trend of the Marginal Syncline, and the facies belts in Triassic sediments (Wellman, 1952 and 1956). The contours show the trend of the anomaly, but being based on strongly smoothed values, lack the detail required for correlation with geological structure. The magnetic results are more readily correlated with geology by changing Gerard and Lawrie's total-value-profiles to anomaly-profiles, and then plotting them on a map (Text-fig. 2). Of the eight profiles shown, only four—the two northern and the two southern—can be related to the pre-Tertiary structure through surface outcrops, the remainder being either over the sea or over Tertiary covering beds.

III. Relation of Magnetic Profiles to Geology

Upper Palaeozoic rocks are well exposed at the north end of the South Island, and the two southern profiles (G and H of Text-fig. 2) will be considered first. Profile G extends east from Moores Valley near Wellington to Puponga at the northern extremity of the South Island. The first 40 miles is probably entirely over greywacke of lower Mesozoic age, the next 10 miles is over the low-rank Marlborough Schist—a slightly metamorphosed greywacke—and the remaining 15 miles to the eastern shore of D'Urville Island over Upper Palaeozoic greywacke. For the whole of this distance of 65 miles the magnetic profile is undisturbed. Greywacke has the very low magnetic susceptibility of

30×10^{-6} C.G.S. units (Cullington, 1954, p. 17) and observations on greywacke are typically undisturbed.

The relation of the strongly disturbed next 60 miles of the profile to the geology is shown by Text-fig. 3. The first peak is caused by the well-known belt of serpentine and basic volcanics through the eastern side of D'Urville Island. The magnetic disturbance is known at ground

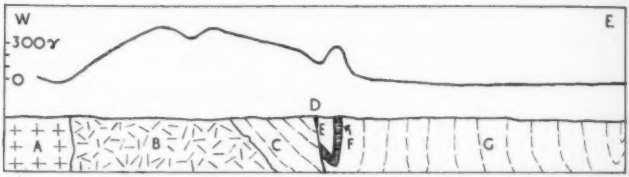


TEXT-FIG. 2.—Sketch map showing airborne magnetometer anomaly profiles and a tentative correlation of anomalies (see Text-fig. 1 for relation to major structural features).

S. Serpentine at Wairere.
K. Kawhia Syncline.

Vertical scale gives total magnetic force in gammas.

level by variations in compass bearings of over 10 degrees across the west margin of the serpentine. The saddle to the east of the magnetic peak reflects a tight syncline in non-magnetic limestone and siltstone. On the west side of D'Urville Island basic lavas and agglomerates of the Brook Street volcanics dip east into the fault and cause the magnetic rise west of the saddle. The remaining part of the profile is over the sea and the geology shown on Text-fig. 3 is inferred from that south of Nelson. There the Brook Street volcanics pass west into a wide belt of coarse-grained acidic to basic rocks known as the Rotoroa Gneiss as seen in



TEXT-FIG. 3.—Diagrammatic geological section along profile " G " through D'Urville Island with magnetic values above. Vertical scale gives total magnetic force in gammas.

A. Separation Point Granite.
B. Rotoroa Gneiss.
C. Brook Street Volcanics.
D. Infaulted strip of Triassic conglomerate.
E. Syncline in Maitai slates and limestone.
F. Upper Te Anau (includes serpentine and volcanics).
G. Upper Paleozoic greywacke grading east into Marlborough Schist.

Length of section 110 miles.

the 1958 map of New Zealand (N.Z. Geological Survey). Two of the largest total force anomalies (740 and 510 gammas) observed during the ground magnetic survey of New Zealand are on the Rotoroa Gneiss at Foxhill and Lake Station (Cullington, 1954 : total force map) and it is inferred that the very high values west of D'Urville Island are caused by the northward extension of the gneiss. The lower values still farther to the west are probably caused by the northward extension of the granite of Separation Point, the promontory 10 miles south of the flight line.

Profile H extends west only to the City of Nelson, and closely matches the corresponding part of profile G. The undisturbed eastern part is over greywacke and schist. The sharp peak is due to the belt of Upper Palaeozoic lavas and serpentine, the saddle to the east is due to the syncline of Upper Palaeozoic sediments—wider here than at D'Urville Island—and the peak at the west end of the profile is due to the Brook Street volcanics.

The close agreement between the magnetic anomalies and the

Upper Palaeozoic structure of the north end of the South Island makes a similar agreement likely in the North Island. But there, with the possible exception of serpentine at Wairere near Te Kuiti (Fleming, 1948), the Upper Palaeozoic sediments and igneous rocks are hidden beneath Tertiary and lower Mesozoic sediments and the Upper Palaeozoic structure cannot be determined directly. If the Upper Palaeozoic igneous rocks are assumed to extend over the region as a folded sheet, then the magnetic highs should correspond to anticlines and the lows to synclines. The probable conformity between the Upper Palaeozoics and the lower Mesozoics makes it likely that folds in the Upper Palaeozoic volcanics continue up through the lower Mesozoics to the surface. Unless the axial planes are vertical, the crests of the folds in the Upper

TEXT-FIG. 4.—Diagrammatic section along profile B through Kawhia Harbour with magnetic values above. Vertical scale gives total magnetic force in gammas. Post-Jurassic rocks not shown.

A. Upper Jurassic.
B. Lower Jurassic.
C. Triassic.
D. Hypothetical Upper Palaeozoics with lava and serpentine.
E. Geosynclinal Triassic.
F. Geosynclinal Jurassic.

Length of section 90 miles.

Palaeozoic volcanics will not be directly below those at the surface, but will be displaced in the direction of the dip of the axial plane of the folds. The amount of displacement will depend on the angle of dip of the axial plane and on the stratigraphic interval between the magnetic rocks and those exposed at the surface.

The Kawhia Syncline is the only well located Mesozoic structure (Marwick, 1947). It has a steep east and a gentle west flank, the axial plane dipping east at about 45 degrees (Text-fig. 4). The position of the synclinal axis at the surface relative to the line of magnetic lows on the profiles is shown by Text-fig. 2. The lows are displaced four miles east from the axis at the surface, the direction being in agreement with the direction of dip of the axial plane, and the amount with the possible thickness of the strata between the syncline in the magnetic rocks and the syncline at the surface.

Triassic beds on the east flank of the Kawhia Syncline lie between the Jurassic belt in the core of the syncline and a Jurassic belt farther to the east and must represent an anticlinal high (Fleming, 1953),

but because of extensive cover the position and nature of the axis is poorly defined. It is suggested that the well-defined magnetic high represents this anticlinal axis in the Upper Palaeozoic igneous rocks and that it will be a short distance west at the surface. The isolated serpentine outcrop at Wairere near Te Kuiti (Fleming, 1948), marked by " S " on Text-fig. 2, is not far west from the line of magnetic highs is probably an outcrop of the igneous rock causing the high magnetic values.

IV. Conclusions

Profiles G and H correlate closely with the exposed geology and serve as controls. Profile F has few features in common with those to the south or north and cannot be correlated with them, its most important feature being that the well-defined eastern limit to the magnetic disturbances lines up with the eastern limits on the lines south and north. For the series of profiles E and A a tentative correlation of the magnetic features is possible. The magnetic high, considered to represent an anticline in the Upper Palaeozoic igneous rocks, can be traced through the five profiles. The low to the west in profiles B and C probably represents the Kawhia Syncline at depth, but is not well shown by A or D.

The aeromagnetic lines already flown have provided invaluable information on structural trends that could hardly be obtained by any other method. Knowledge of New Zealand structure would be greatly benefited if more observations were made over Cook Strait, and observations extended south over the other belts of Upper Palaeozoic igneous rocks in the South Island and their probable seaward extension (Text-fig. 1).

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Victoria University of Wellington,
New Zealand.

The Mt. Barney Central Complex, S.E. Queensland

By P. J. Stephenson

Abstract

The general geology of the Mt. Barney Central Complex is described. This intrusive complex shows close similarity in type to the classic Tertiary ring complexes of Western Scotland and Northern Ireland. In this discussion of the complex, the similarities shown by ring dykes, cone sheets, a forcefully intruded granophyre boss, and a closed ring fault are indicated.

I. Introduction

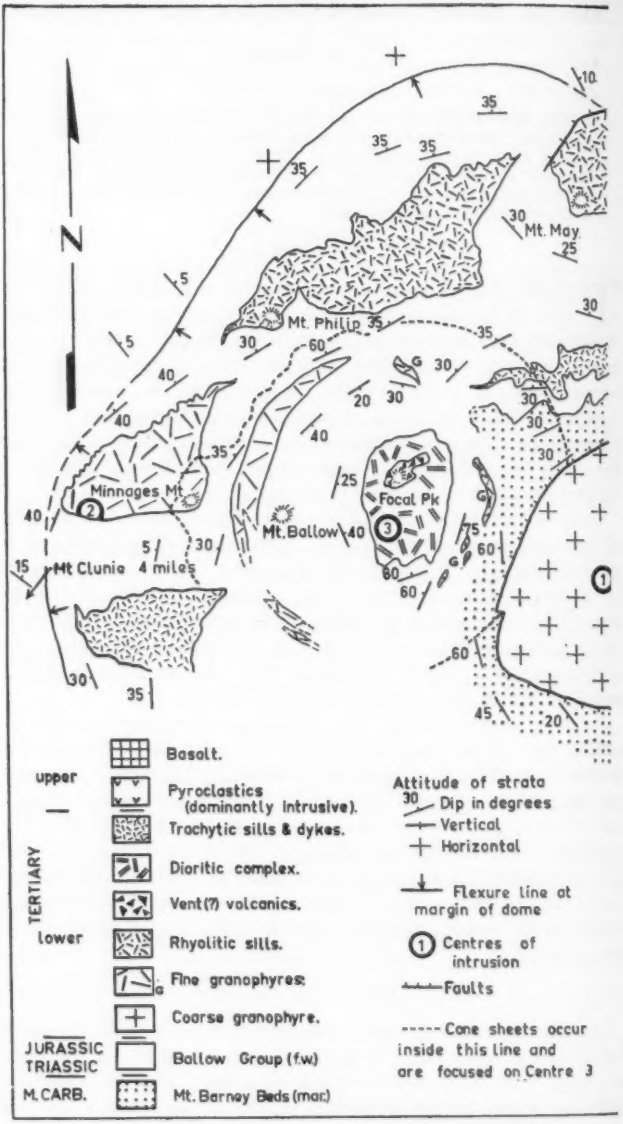
THE Mt. Barney area lies some seventy miles south of Brisbane and the igneous complex with which this paper deals is situated astride the Queensland–New South Wales border some sixty miles inland from the coast. The complex forms mountainous country whose peaks rise to about 4,400 feet and which is clothed in open eucalyptus forest or thick sub-tropical rain-forest that hampers investigation. The area occupied by the complex itself is not inhabited, but access to it is provided by roads serving the surrounding farmlands which have a general elevation of about 800 feet.

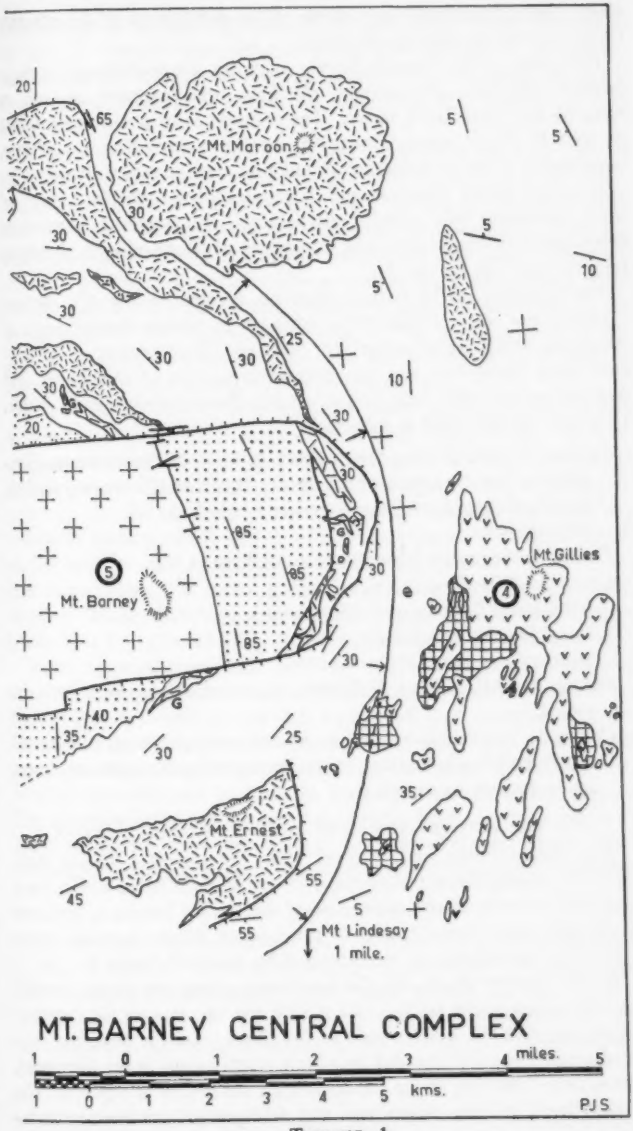
Early in 1950 it was noticed that aerial photographs recently taken of the Mt. Barney area revealed some unusual concentric topography suggestive of possible ring structures. These observations led to an investigation of the area over the years 1950–54; during this period the existence of ring structures was confirmed and the character of the igneous complex was more fully explored. A general account of the complex is given in the present paper and certain aspects of its petrology will be dealt with in more detail elsewhere.

Previous workers in the Mt. Barney area investigated only the eastern side of the complex and its true nature was therefore not apparent. Richards (1916) described the main intrusive mass of the complex, the Mt. Barney Granophyre of the present writer. Wearne and Woolnough (1911) and Richards, Bryan, and Whitehouse (1932) discovered near Mt. Barney inliers of marine Carboniferous rocks; the Carboniferous is not in general exposed in south-east Queensland and, as will be shown later, its appearance is connected with the formation of the igneous complex. Various types of volcanic rock were also recorded.

II. Outline of Geology

The country rocks of the Mt. Barney complex consist of marine Carboniferous rocks unconformably covered by freshwater Mesozoic sediments with some inter-stratified volcanics. The Carboniferous rocks occur only in the immediate vicinity of the igneous complex


N
Mt. Philip
Mt. May
Minnages Mt
Mt. Ballow
Focal Pk
Mt Clunie 4 miles
Basalt.
Pyroclastics (dominantly intrusive).
Trachytic sills & dykes.
Dioritic complex.
Vent(?) volcanics.
Rhyolitic sills.
Fine granophyres.
Coarse granophyre.
Ballow Group (f.w)
Mt. Barney Beds (mar.)
upper
lower
TERTIARY
JURASSIC
TRIASSIC
M.CARB.
Attitude of strata
Dip in degrees
Vertical
Horizontal
Flexure line at margin of dome
Centres of intrusion
Faults
Cone sheets occur inside this line and are focused on Centre 3


Mt. Maroon
Mt. Barney
Mt. Gillies
Mt. Ernest
Mt Lindesay
1 mile.
MT. BARNEY CENTRAL COMPLEX
miles.
kms.
PJS

TEXT-FIG. 1.

where they are exposed as a result of the doming-up of the country rocks and the up-faulting of a plug.

The rocks of the Mount Barney complex are varied and include diorites, dolerites, granophyres, rhyolites, and trachytes as well as basic to acid pyroclastic rocks. The complex is of particular interest in that it shows special intrusive structures including ring-dykes, cone-sheets, a broad dome associated with the forceful intrusion of a boss, and a closed ring-fault with a strong central upthrow. Many of these structures are closely comparable with intrusive structures described from the central ring complexes of Scotland and Northern Ireland (see Text-fig. fig. 1).

The mapping of the various elements composing the Mt. Barney Complex has made it possible to establish in outline the sequence of events by which the complex was built up. Five centres of activity have been recognized and the descriptive section of this paper will deal in turn with the rocks and structures associated with each centre. They may be tabulated as follows:—

Centre 1: Mt. Barney Granophyre (a boss-like intrusion responsible for the formation of the Barney Dome in the country rocks).
Barrier Granophyre (an incomplete ring-dyke).
Rhyolite sills.
Centre 2: Minnages Mountain Granophyre (a sill).
Centre 3: The Central Volcanics.
Montserrat Granophyres (ring-dykes).
Focal Diorite Complex.
Ballow Cone-sheets.
Centre 4: Mt. Gillies Volcanics (vent intrusions and extrusive rocks).
Centre 5: No visible igneous rocks are associated with this centre, but its position is marked by the Barney Ring Fracture, enclosing an up-faulted central plug.

III. The Country Rocks of the Mt. Barney Complex

The *Carboniferous rocks*, which make up the *Mt. Barney Beds*, have two developments; they occur in the Western Inlier which has a crescentic outcrop in the inner part of the Barney Dome, a structure associated with Centre 1, and in the Eastern Block, portion of the upthrust plug enclosed by the Barney Ring Fault of Centre 5.

In the Eastern Block, the Carboniferous strata are nearly vertical and the oldest rocks are in contact with the Mt. Barney Granophyre which occupies the western part of the block. On the east, the Carboniferous rocks are bounded by a fault which brings in the Mesozoic. More than 7,000 feet of the Carboniferous succession is exposed. The lowest beds include limestones and feldspathic sandstones, meta-

morphosed to fine-grained calc-silicate rocks and biotite hornfelses by the Mt. Barney Granophyre. These are followed by feldspathic sandstones and shales with interbedded tuffs which pass up into some 6,000 feet of feldspathic sandstones, sandy and silty shales, and thin conglomerates.

In the Western Inlier, the oldest rocks appear at the inner side of the crescentic outcrop and the strata dip outwards at angles of 25° to 60°. The inner margin of the outcrop is made by the Barney Ring Fault and the outer boundary by the Mesozoic unconformity. Between 2,000 and 3,000 feet of feldspathic sandstones and sandy shales are exposed. A close correlation with the rocks of the Eastern Block is not possible but fossil remains and lithology suggest that the rocks of the Western Inlier are comparable in age to the youngest part of the succession in the Eastern Block. The full development of the Mt. Barney Beds is thus between 7,000 and 10,000 feet.

Freshwater *Mesozoic sediments* are widely developed over much of south-east Queensland and northern New South Wales. Jurassic rocks are exposed in the region surrounding the Mt. Barney Complex and Triassic sediments are brought to view above the Carboniferous inlier on the flanks of the Barney Dome. In general, the rocks dip outwards from the centre of the dome and flatten off rather abruptly at the margin of this structure. Beyond the area disturbed by the igneous complex, the rocks display a broad warping associated with faulting. Within the complex, anomalous dips have been observed within some of the ring structures; their significance will be discussed later.

A succession of more than 7,000 feet of Mesozoic rocks, making up the *Ballow Group*, has been established. On the northern side of the Barney Dome, conglomerates (the Yamahra Conglomerates), followed by a variable thickness of basic volcanic rocks, rest on the basal unconformity. They are succeeded by quartzose sandstones with interbedded siltstones and shales and some calcareous sandstones. The conglomerates and the lower members of the overlying sediments are considered to be Upper Triassic, from fossil plant evidence, while the middle and upper parts of the group contain Jurassic plant remains.

IV. The Igneous Complex

The petrography and structure of the most important components of the Mt. Barney Igneous Complex are described in this section. Certain topics of more general petrological interest, such as the composition and optics of the feldspars, and the development of and significance of granophyric textures, will be discussed in more detail elsewhere.

(*a*) *Centre 1*

(i) The Mt. Barney Granophyre: This body of coarse granophyre

composes the mass of Mt. Barney itself and is the nucleus of the complex. Its outcrop is broadly elliptical in outline, but much of its contact is faulted. The original intrusive contact is preserved only in the east, where the granophyre invades Carboniferous sediments of the Eastern Block; this contact is broadly conformable and dips very steeply away from the granophyre. On the west, north, and south sides the contact of the granophyre is marked by the Barney Ring Fault which developed from a later centre.

The granophyre lies at the centre of the Barney Dome, in which the country rocks show quaquaversal dips away from the granophyre itself. This dome is a local structure, and has been formed by the forceful intrusion of the Mt. Barney Granophyre, in a very similar way to the Arran Dome of Scotland.

Petrographically, the Mt. Barney Granophyre is very constant in character, and the only marked variations in it occur at contacts. Feldspar phenocrysts (sanidine cryptoperthite) occur up to 5 mm. in diameter and, accompanied by corroded quartz phenocrysts, fine plates of biotite, and granular iron oxide, are set in a moderately coarse granophyric matrix. Near contacts, very local modifications of texture and mineralogy are produced.

(ii) The Barrier Granophyre: This intrusion of fine granophyre forms a broken arc, to the east of the Mt. Barney Granophyre. The terminations of this arc are faulted by the Barney Ring Fault, but fine granophyre reappears outside the fault further west, both to the north and to the south; these two western masses are considered to be parts of the same intrusion, which would thus originally have been semi-circular in form. The contacts all appear to be vertical or steeply-dipping and the intrusion may therefore be regarded as an incomplete ring-dyke.

The pattern of outcrops is discontinuous, the granophyre appearing as a succession of aligned, dyke-like masses. The intrusions do not coincide with a recognized fault along the whole of their course, although for some distance they follow the faulted junction of the Carboniferous and Mesozoic sediments within the Barney Ring Fault; the granophyre, however, sends a radial arm across this fault and therefore appears to be younger than it.

The mode of occurrence of the Barrier Granophyre, in discontinous outcrops, suggests an arcuate line of vents rather than a continuous intrusion. There are, however, no associated pyroclastic rocks and the texture of the granophyre itself suggests that it crystallized as such and does not represent altered pyroclastics. It is possible that early vent action cleared the way for the emplacement of magma producing granophyre.

The Barrier Granophyre is petrographically similar to the Mt.

Barney Granophyre, but the granophyric intergrowth is much finer grained.

(iii) Rhyolite Sills: Several mountains in the region are made up of rhyolite sills which are broadly conformable and which intrude the Mesozoic rocks on the flank of the Barney Dome. Four main sills have been mapped, on Mt. May, Mt. Philip, Mt. Maroon, and Mt. Ernest; the Mt. May and Mt. Philip sills may originally have been continuous. The thicknesses of the sills are very variable and in several places the intrusions can be seen to be wedge-shaped. The Mt. Maroon sill has a thickness of more than 1,200 feet. No dyke-feeders are exposed but the direction of observed thinning of the sills suggests that the feeders are concealed beneath the present outcrops.

The rhyolites are buff-coloured rocks composed of phenocrysts of sanidine cryptoperthite and quartz, set in a finely crystalline groundmass which is granophyric in coarser specimens.

(*b*) *Centre 2*

The Minnages Mt. Granophyre: Minnages Mt., on the west side of Mt. Barney, is made up of a granophyre of a different type from the Mt. Barney and Barrier granophyres. This rock is composed of plagioclase and alkali feldspar, quartz, hornblende, pyroxene, biotite, iron oxides, and accessory minerals. The pyroxene is unusual in being pigeonite and the feldspars have cloudy microperthitic alkali feldspar mantles and strongly zoned andesine-oligoclase centres. Granophyric intergrowth occurs between quartz and alkali feldspar, forming the matrix.

The Minnages Mt. Granophyre forms a tabular body with a general north-westerly dip of 25° to 35°, which is parallel to the dip of the adjacent Mesozoic sediments. At most of the exposed contacts it is conformable, but at one point near the south-western corner of the outcrop there is a steeply dipping contact, dipping 75° S, at this point. This steep attitude suggests the feeder of the intrusion may be concealed at or near this locality.

(*c*) *Centre 3*

The third centre of activity in the complex is the most diverse. Four individual units occur and are of widely different character, with both ring dykes and cone-sheets represented. The arrangement of the intrusions around the centre (near Focal Peak), is reflected in a most striking manner in the topography, giving rise to a strong concentric and radial pattern.

(i) The Central Volcanics: On the upper slopes of Focal Peak there occur extrusive and pyroclastic rocks of basic and sub-acid character, which appear to form a capping on the Focal Diorite Complex. These

volcanic rocks show evidence of contact metamorphism, however, due to the diorite complex, and must therefore be earlier than this intrusion. There is no evidence available in the outcrops to show that they are layered surface extrusions and their occurrence in a structurally deep part of the complex suggests they might be early vent volcanics. No really coarse pyroclastic rocks have, however, been found.

On Focal Peak itself the volcanics are made up dominantly of trachytic agglomerates but andesitic and basaltic lavas appear in the eastern part of the outcrop and andesitic and basaltic pyroclastics occur on the lower slopes to the north-west. Fragments in the agglomerates include porphyritic rhyolites of a type similar to the rocks making up the rhyolite sills of Mt. May and other localities already described.

(ii) The Montserrat Granophyres: Fine-grained granophyres make a discontinuous oval ring with a diameter of some 2½ miles, circling Focal Peak. They appear to form a number of arcuate dykes, separated by screens of country rocks, and their steep attitude indicates they are of ring-dyke type.

The structure of the country rocks which are invaded by the granophyres show some features of special interest. Outside the ring the rocks dip at moderate angles towards the north-west in directions which are controlled by their position in the flanks of the Barney Dome. Inside the granophyre arcs, the dips of the country-rocks are directed inward towards Focal Peak, forming a basin-like structure. It seems probable that this structure was formed as a result of subsidence of a central mass of country rocks.

The Montserrat Granophyres are fine-grained porphyritic rocks which are in most respects petrographically similar to the Minnages Mt. Granophyre. They are, however, devoid of pigeonite, and hornblende is the characteristic ferromagnesian mineral. The phenocrysts are generally zoned plagioclase, and two generations of this mineral can be recognized. The amphibole, often altered to chlorite, in some cases also appears in two generations. The matrix of the rock is made up of fine granophyric intergrowth of quartz and turbid alkali feldspar.

(iii) The Focal Diorite Complex: The country surrounding Focal Peak is made up of a variable complex of dioritic rocks, which are poorly exposed in most places. The shape of the body cannot be precisely determined, but it forms an oval area of about 1 by 1½ miles. The actual contact, with Mesozoic sediments, has been seen only in two localities and in these it appears to be steep, and the outline of the complex as a whole also suggests that its contacts are steep. The Central Volcanics apparently represent part of the original roof and this fact suggests that the diorites are exposed at a relatively shallow level. The complex is surrounded by a zone of closely-jointed and

fractured country-rocks and this fracturing is probably related to the emplacement of the diorites. However, the dips of the country-rocks are generally directed towards the complex and provide no evidence of forceful emplacement.

The diorites of the complex are variable in appearance. The rock-types include even-grained diorites, " patchy " mottled dioritic rocks and more feldspathic rocks, and are composed of zoned feldspar, diopsidic augite, chlorite, ilmenite, quartz, and accessory minerals. The variable dioritic rocks are gradational with one another. At the contacts with quartzose sandstones, a narrow layer of granophyre, less than an inch wide, is formed from the sediment.

(iv) The Ballow Cone–sheets: A series of dolerite intrusions are developed around Focal Peak over a circular area approximately four miles in diameter. Most of these intrusions are from 3 to 5 feet thick. Their strikes are tangential and they dip consistently towards Focal Peak at angles of from 25 to 60 degrees. Adjacent dykes are sub-parallel and are separated by screens of country-rock. These intrusions are cone-sheets and are remarkably similar to the basic cone-sheets of the central complexes of Scotland and Northern Ireland.

The Ballow Cone-sheets largely control the concentric topography surrounding Focal Peak. Their strike determines the trend of ridges and streams in a manner well displayed on Double Peak.

The cone-sheets intrude all of the other rocks associated with the Focal Peak centre of intrusion and therefore represent the youngest unit of this centre. They also intrude the Minnages Mt. Granophyre, but they are not seen to intrude the Mt. Barney Granophyre; it is due to later faulting along the Barney Ring Fracture that the cone-sheets cannot be seen to intrude this granophyre.

The rocks of the cone-sheets are dark-coloured dolerites. The material from the centre of dykes is moderately coarse, with feldspars and pyroxenes up to 2 mm. in length. Smaller cone-sheets are fine grained and contacts are generally quite aphanitic. Veining of the border zone of the intrusions is very common; these veins are mostly parallel to the contact and are quartzo-feldspathic in type. The dolerites are generally non-porphyritic and are composed of zoned acid labradorite, titaniferous diopsidic pyroxene, chlorite, a little turbid alkali feldspar, ilmenite, and accessory minerals.

Xenoliths of even-grained diorite are common and are petrographically similar to the diorites of the Focal Diorite Complex.

(*d*) *Centre 4*

The Mt. Gillies Volcanics: These volcanic rocks make up the rugged area of Mt. Gillies on the eastern side of the Mt. Barney Complex. From their nature and disposition the majority of these rocks appear

to be intrusive vent rocks, including such types as fluidal rhyolites, rhyolitic tuffs, and agglomerates and subordinate basic agglomerates. There is, however, a restricted locality where bedded rhyolitic pyroclastics are found, on the western side of Mt. Gillies; the significance of these extrusive rocks is uncertain, but they may represent the remnants of crater-filling deposits.

The Mt. Gillies Volcanics tend to outcrop in elongated belts; fluidal structures, both in rhyolites and in pyroclastic rocks, are steeply inclined and their distribution confirms that these elongated outcrops are vents. The vents intrude Mesozoic rocks and are also intrusive into the lowest members of the neighbouring volcanic succession on Mt. Lindesay. There is evidence that these Mt. Lindesay Lavas and therefore the Mt. Gillies Volcanics intruding them, are considerably younger than the main plutonic bodies of the Mt. Barney Complex. Structurally, however, the intrusives on Mt. Gillies are a continuation of the igneous activity of the complex, as will be indicated in the concluding section.

(*e*) *Centre 5*

No visible igneous rocks are associated with this centre, but the activity of the centre is marked by the Barney Ring Fracture which completely encloses an up-faulted plug. Its nature and significance are described in the final section of this account.

V. Structural Geology of the Complex

Several igneous structural features are noteworthy in their close similarity to comparable structures in the British Tertiary Central Complexes.

The Barney Dome.—This feature has been referred to above, in the discussions on the country-rocks and also on the Mt. Barney Granophyre. It may be compared with the well-known Arran Dome.

Each of these domes has been produced by the forceful intrusion of a boss of acid plutonic rock; the Mt. Barney Granophyre is closely similar to the Arran Granite, not only in its environment but also in its composition. Even the size of the domes is closely similar, the Arran Dome being approximately 7 miles in radius (towards the south), the Barney Dome 5 to 6 miles in radius. A further similarity is shown in the fact that both structures were formed very early in the intrusive history of their respective complexes.

Ring-Dykes.—The ring-dykes in the Mt. Barney Complex are comparable in type with examples from the British Tertiary, but they are considerably smaller and simpler.

The inclination of contacts is not certain, since contacts are rare. Suggestions indicate that the Barrier Granophyre and the Mont-

serrat Granophyres are approximately vertical in their attitudes. There is no evidence of central subsidence in the case of the Barrier Granophyre, but as described earlier, the Montserrat Granophyres do appear to have been emplaced in this manner.

Cone-Sheets.—The Ballow Cone-sheets are remarkably similar to examples in Ardnamurchan, Skye, and Mull. Their development is smaller, however, and their depth of focus is significantly shallower. Whereas examples in Scotland have foci lying at depths of close to 3 miles below the existing land surface, those at Mt. Barney lie at depths of 1 to 1½ miles. The difference of depth can scarcely be due to a difference in depths of erosion (if anything the Mt. Barney Complex is exposed at a shallower depth) and the Ballow Cone-sheets must simply have been intruded from shallower depths.

The Barney Ring Fracture.—This closed circular fault marks the boundary of the Mt. Barney Granophyre on three sides and is continued eastwards to enclose the Eastern Block of country-rocks which includes a block of vertically inclined Carboniferous sediments and also a group of Mesozoic sediments. No igneous rocks are associated with this fault, but it is considered to be the result of concealed forceful intrusion.

As an example of igneous tectonics it is comparable with the circular fault on the island of Rhum, described by Bailey (1944). Like this fault on Rhum, the Barney Ring Fracture is approximately vertical. The diameter of the fault on Rhum, 5 miles, may be compared with that on Mt. Barney, 5 miles by 3 miles. Bailey has estimated that uplift inside the fault at Rhum was as much as 8,000 feet at some points. Estimation of the fault at Mt. Barney is difficult owing to the complexity of the structures of the country-rocks inside the fault, and the inability to correlate these structures with any outside the fault (the belt of vertical Carboniferous, for example). The discrepancy of structure across the fault suggests large displacement and indirect estimates utilizing the possible depth of erosion suggest that the Barney Ring Fracture was at least of the same order of displacement as the one on Rhum.

The Relation of the Intrusion Centres.—The centres of intrusion, and the igneous rocks identifying them, have been described above, in the probable order of emplacement. Evidence for establishing a sequence of intrusion is indirect in many cases and for the intrusion of centre 2, the Minnages Mt. Granophyre, there is no evidence for its age relative to the rocks of centre 1. By analogy with the Scottish examples, it has been assumed in determining the sequence of intrusion that igneous activity was confined to successive centres—once a new centre had been initiated, activity did not revert to an old one.

If the location of the centres of intrusion are each considered on the

map (Text-fig. 1), it is noticeable that they are broadly co-linear. This aligning of the centres must express a structural control on the localization of intrusion, but no indication of such a feature in the exposed country-rocks has been recognized. The alignment of the main mass of vent rocks on Mt. Gillies with the centres further west vindicates the inclusion of these youngest rocks as members in the Central Complex.

The Environment of the Complex and its Age.—The complex has been emplaced into Palaeozoic and Mesozoic country-rocks in what has been a regionally stable part of the Australian continent since late Palaeozoic to early Mesozoic times when the last orogenic movements took place.

The Mt. Barney region is situated in a Tertiary volcanic province in south-eastern Queensland. While the igneous complex described here lies adjacent to thick extrusive basic volcanics (on Mt. Lindesay and Mt. Cluney), it does not show any relation to these, and since boulders of Mt. Barney Granophyre are found in boulder beds interbedded with these basic volcanics further east, the intrusive complex must be older than these lavas. The lavas are considered to be of Tertiary age, and if the preliminary results of spore studies (occurring in interbedded shales) are to be accepted, then the age of the complex can be defined as post-Triassic and pre-Eocene-Oligocene.

Acknowledgments

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Department of Geology,
University of Queensland,
St. Lucia, Brisbane.

Steeply Plunging Recumbent Folds.

By KSHITINDRAMOHAN NAHA

ABSTRACT

The definition of a recumbent fold as one with subhorizontal axial surface is beset with difficulties when recumbent folds plunge. Recumbent folds are redefined here as folds whose axial planes give *subhorizontal trace in profile plane*, with the fold-axis plunging at any angle up to 80°. A neutral to slightly antiformal recumbent fold half mile across and plunging 75° is described from the Pre-Cambrian metamorphic terrain in Eastern India.

I. GEOMETRY OF PLUNGING RECUMBENT FOLDS

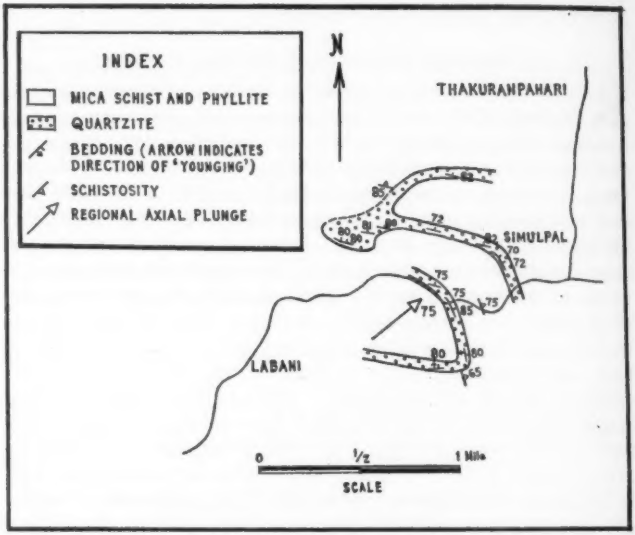
RECUMBENT folds are defined in textbooks of structural geology (Billings, 1942, p. 40) as folds whose axial surfaces are essentially horizontal (dipping up to, say, 20°). This definition, suited for recumbent folds with horizontal fold-axes, does not suit those with plunging axes; and many of the recumbent folds described in literature have their axes plunging at any angle up to 30° (Bailey and McCallien, 1937; McIntyre, 1951, 1952; Stockwell, 1950). When folds are plunging down the dip of their axial surfaces, irrespective of the amount of plunge of fold-axes and dip of axial planes (excepting the limiting case of 90° plunge and dip respectively) the profiles drawn at right angles to the fold-axes (tectonic profiles, Wegmann, 1929; McIntyre, 1951) will have a horizontal trace of the axial planes, thus retaining the characteristic profile of recumbent folds. Recently, such folds with axial plunge down the dip of axial surfaces and with sidewise closure have been termed " reclined folds " by Fleuty (discussion of paper by Johnson, 1957, p. 268). It should, however, be emphasized that these folds possess all the important geometrical peculiarities of the " orthodox " recumbent folds: (*a*) In flat ground, the direction of plunge in one-half of the folds is opposite to the direction of " younging ", with consequent repetition of the same bed along the plunge. (*b*) The dips of beds at the hinge of the folds are maximum (vertical) as in " orthodox " recumbent folds, in contrast with non-recumbent folds in which the dip is minimum at the hinge.

The ambiguity vanishes if recumbent folds are defined as folds *whose axial surfaces give a subhorizontal trace in the tectonic profile.* It follows that if the fold-axis plunges nearly down the dip of axial planes, folds with dip of limbs and axial planes nowhere less than 89° will still be called recumbent, with the fold-axis plunging 89°. It is however obvious that, as the axial plunge approaches 90°, the axial trend becomes less and less significant, and correspondingly the strike of the profile plane becomes less and less suited as a reference direction. Taking 80° as an arbitrary limit (with the axial trend still significant),

therefore, a recumbent fold is defined as one whose axial plane gives a subhorizontal trace (up to 20° pitch) in the profile plane, with the axial plunge at any angle up to 80°.

II. Folds Near Simulpal

The steeply plunging recumbent fold to be described occurs west of Simulpal (86° 42′: 22° 36·5′) near the border of Midnapore and Singhbhum districts in the Pre-Cambrian terrain of Eastern India.

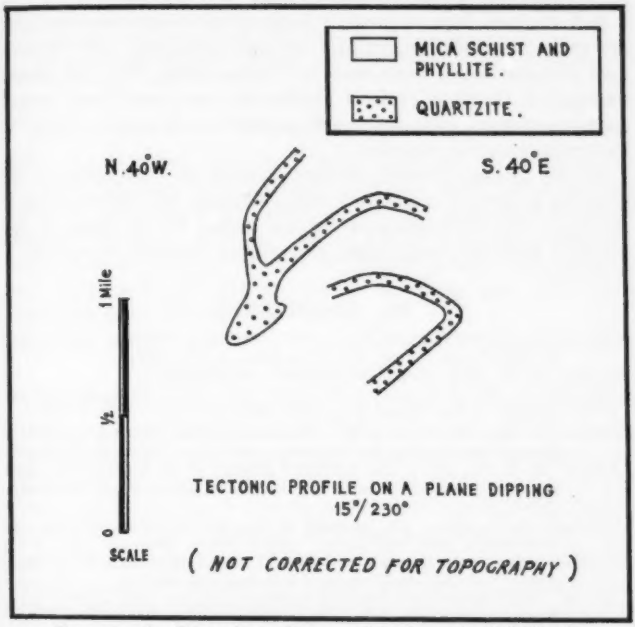


Text-fig. 1.—Map of the area around Simulpal.

Here, in a region of N.E.- and S.W.-closing folds, two quartzite bands around Simulpal show W.N.W.- and E.S.E.-closures (Dunn and Dey, 1942, pl. 24). Owing to this peculiarity in outcrop pattern the two quartzite bands were studied in detail in the background of the regional setting.

Bedding dips in the quartzites are nowhere less than 70°, and generally not less than 75°, becoming steeper with swing in strike towards N.E. near the hinge (Text-fig. 1). The lowest dips are always towards N.E., suggesting that the major fold plunges towards N.E. (Wegmann, 1929). Poles of bedding planes (data exclusively from the two quartzite bands) fall in a well-formed girdle, with the pole of the girdle plunging 75° towards 50°, indicating that the region is tectonically homogeneous,

with the fold-axis plunging 75° towards N. 50° E. Rare small folds in the quartzites also plunge steeply towards north-east. Axial plane schistosity in the associated phyllites and phyllitic mica schists changes in strike from N.N.W. south-west of Simulpal to W.N.W. west of the village, dipping steeply towards north-east. The dip of the axial plane is thus towards the direction of axial plunge.



TEXT-FIG. 2.—Tectonic profile normal to the regional fold-axis.

A tectonic profile has been constructed from the map (Text-fig. 2) by the axial projection of the two quartzite horizons on the profile plane (normal to the fold-axis) which dips 15° towards S. 50° W. (cf. Wegmann, 1929; McIntyre, 1951). For this construction it is assumed that the fold style remains the same along the axial plunge, as the fold-axis is by definition the direction of maximum continuity in space (Clark and McIntyre, 1951). No correction for topography was deemed necessary, as the fold plunge is very steep, and in comparison with that the topography is almost flat.

The profile shows that the southern quartzite (the quartzite in the lower tectonic level in the profile) is antiformal to almost neutral

closing towards south-east, with the trace of the axial plane at an angle of up to 15° with the horizontal. The upper level quartzite is synformal, becoming antiformal nearly above the hinge of the lower level quartzite. The southern quartzite shows indistinct current bedding at two points as shown in the map (Text-fig. 1). Taking these readings as significant, the southern quartzite is a neutral to antiformal recumbent syncline (with younger beds in core).

This structure forms a part of a larger tectonic unit where the folds are upright to slightly overturned towards north-west, with the fold-axes plunging steeply north-eastward (Naha, 1956). The fold around Simulpal is therefore of local significance, and must have formed contemporaneous with the nearly upright north-easterly folds, by local recumbency at the nose.

Mr. K. L. Das provided excellent facilities for field work and his generous hospitality is gratefully acknowledged. Mr. Subir Ghosh of Jadavpur University accompanied the author to the field on one occasion, discussed many points, and critically read the manuscript.

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INDIAN INSTITUTE OF TECHNOLOGY,
KHARAGPUR.

Chess-board Albite from New Brunswick, Canada

By John Starkey

(PLATE III)

Abstract

Chess-board albite is described and theories on its origin outlined. A new occurrence from New Brunswick, Canada, is noted and certain features of the chess-board albite, discussed which fit the theory that it results from two processes, namely metasomatism of potash feldspar to soda feldspar, and deformation. This hypothesis is applied to British localities.

Introduction

IT is just over fifty years since Becke (1906, pp. 124–125) first described Schachbrettalbit (chess-board albite); since then it has been mentioned sporadically in geological literature, particularly that pertaining to the Alps. Although Flett (in Peach, *et al.*) recorded chess-board albite as long ago as 1911, it appears to have been generally overlooked by British authors; this is surprising since chess-board albite does occur in many British rocks.

Description of Chess-Board Albite

Chess-board albite is distinguished by a characteristic development of twinning on the albite law resulting in a complex pattern of short alternating twin lamellae which, unlike usual albite twin lamellae, do not pass through the entire crystal. Consequently when seen in thin section, cut perpendicular to (010), under crossed nicols, a chess-board pattern results when one set of twins is extinguished (Plate III, fig. 1 and fig. 3). As the plane of the thin section approaches (010) the chess-board pattern becomes less distinct until finally, when cut parallel to (010), the albite has a patchy appearance resembling strained extinction, the edges of the patches being rather indefinite. This is readily appreciated since in all sections sub-perpendicular to (010) the twin lamellae are lenticular then they must be approximately lense-shaped in three dimensions, the longer axes of the lenses being parallel to (010). The overlapping of the feather edges of these lenses produces the indefinite borders to the twins seen in (010) sections.

Review of the Literature

In the opinion of Becke (op. cit.) chess-board albite is of secondary origin, arising by replacement of microcline, a view generally held amongst Alpine geologists. This view was accepted by Grubenmann and Niggli (1924, pp. 401 and 436); they considered that the change

from potash feldspar to chess-board albite occurred in the Katazone of metamorphism.

Anderson (1928, p. 205), in describing Pre-Cambrian granite pegmatites from Norway, also attributed the formation of chess-board albite to replacement of microcline but derived the metasomatizing fluids from the consolidating magma. Gilluly (1933, pp. 68 and 73) found chess-board albite in an albite granite near Sparta, Oregon, U.S.A., and claimed that " the occurrence of this variety of albite, commonly regarded as typical of replacement, lends some additional support to the more direct evidence of a secondary albitisation of these rocks ". Niggli too (1952, pp. 36 and 251) regarded the chess-board pattern as a replacement texture.

Tilley (1919, p. 328) and Battey (1955, p. 73) both ascribe a primary origin to chess-board albite. The former considers that it may result from irregular deposition of albite substance during the growth of a crystal while the latter envisages a process of accumulation and welding together of individual albite laths to give a chess-board pattern.

A fuller treatment is given by Exner (1949) in his discussion of the calc-alkaline igneous rocks of the eastern Hohe Tauern, Austria. He considers that in these rocks chess-board albite arose by two distinct processes. The first is an unmixing of potash-soda feldspar, through *Faserperthit* (thread-perthite) and *Aderperthit* (vein-perthite) to chess-board patterned *Haufenperthit* by a broadening of the exsolution lamellae, while contemporaneous metasomatic replacement of the potash by soda gives true chess-board albite (chess-board patterned soda feldspar). In the second process, according to Exner the more usual, metasomatism is the sole factor; potash-soda feldspar is transformed into pure soda feldspar and the chess-board pattern results from a change in the lattice structure from a high to a low temperature form, the lattice of chess-board albite being related to that of microcline.

The metasomatism in either case is supposed by Exner to be a sudden process, " It works like magic," since he found very few transition members. He regards chess-board albitization, as these processes may be called, as a late phenomenon which occurs during the final stages of consolidation of the igneous rocks.

In view of the conflict of opinion on the origin of chess-board albite it may be of interest to discuss some features of chess-board albite from rocks occurring in Northumberland County, New Brunswick, Canada.

Chess-Board Albite in Porphyries in New Brunswick, Canada

Chess-board albite is common in New Brunswick, Canada, in some porphyries which have been regionally metamorphosed to chlorite and low biotite grade, while some of the associated rocks have been

transformed into albite schists. Where the porphyries are only weakly metamorphosed they contain phenocrysts of perthite and quartz in a fine grained groundmass of quartz, chlorite or biotite and ore. The perthites are twinned according to the Carlsbad and Baveno laws. With increasing metamorphism the groundmass of these porphyries becomes more coarsely crystalline and the quartz phenocrysts disappear, while the perthite is transformed into albite and develops a chess-board pattern. In these chess-board albites the original Carlsbad and Baveno twinning often survives (Plate III).

Transitional stages in the change from potash-soda feldspar to chess-board patterned soda feldspar occur. They are represented in patchy grains, some patches being of chess-board albite and the remainder of a complex nature, consisting of highly irregularly shaped, small grains of albite (approximately ·0015 mm. mean diameter) showing two preferred orientations coinciding with the orientation of the two sets of chess-board twins in the previously mentioned patches, and also of equally irregular and minute potash feldspar grains; unfortunately the grain size is too small to allow optical determinations to be made, and even an X-ray powder photograph failed to reveal the nature of the potash feldspar; all the lines obtained being attributable to albite and quartz, it was impossible to separate all the quartz from the albite.

One important feature which is apparent in the chess-board albite of these rocks is the close connection between the development of the chess-board pattern and zones of deformation. Where cracks occur in the albite crystals the chess-board pattern is best developed alongside them. Often it can be seen that displacement has occurred along these cracks since the composition planes of the growth twins (the Carlsbad and Baveno twins) is offset. Similarly, in other albite crystals, the chess-board pattern is restricted to some zones in the crystal while the intervening " islands " only exhibit a minor development of albite twinning, which, although parallel throughout each individual " island ", may be rotated by up to 2° when traced from one to another; again suggesting that the chess-board pattern has developed along zones of displacement. In addition, the chess-board pattern is often also well developed around the margins of the grains.

In view of this close association of the chess-board pattern with those parts of the crystal which may logically be supposed to have been stressed it is reasonable to assume that this stress is causally related to the chess-board pattern and that the chess-board twinning is a deformation twinning just as the more normal type of polysynthetic twinning (Emmons and Gates, 1953). This is supported by the occurrence of the chess-board albite in the more highly metamorphosed porphyries.

These observations are in accordance with the view held by Dr. G. Voll (to be published) that chess-board albite results from two processes. The first is metasomatism of the potash-soda feldspar into albite, during which the potash feldspar may impart its irregular twin pattern to the replacing albite; this may well be associated with the regional development of albite in the schists of this area and introduction of soda. The second is deformation which results in the chess-board pattern, this being developed in preference to the more normal polysynthetic twinning because the albite has taken over the lattice peculiarities of the potash feldspar. If it should be shown that chess-board albite can develop from original albite, i.e. that it has not been derived by replacement, then an operative mechanism is more difficult to find.

Conclusion

The theory regarding the genesis of chess-board albite outlined above can be applied to described British occurrences.

Chess-board albite has been described from pink intrusive porphyry occurring in the Tayvallich Peninsula, Argyll, (Flett, op. cit., p. 93), from quartz keratophyres of Newton Abbot, Devon (Flett in Ussher, *et al.*, 1911), and from acid volcanics in Caernarvonshire (Hughes, 1917). The " water-clear albite with peculiar streaky twinning recalling at first sight anorthoclase ", described by Cox (1915) from quartz-enstatite diabases in Pembrokeshire may well be chess-board albite. Chess-board albite also occurs in the granite of St. David's Head, Pembrokeshire.

In each of these cases the albite may have arisen by replacement of potassic feldspar in the magmatic or deuteric phases of intrusion, and the chess-board pattern subsequently or contemporaneously by stress. Stress may reasonably be assumed since Emmons (op. cit. and Emmons *et al.*, 1953) considers that both contact between adjacent crystals in a melt and cooling and contraction of an igneous body can set up sufficient stresses to produce polysynthetic twinning, while in the case of the occurrences in the Tayvallich Peninsula and Caernarvonshire there is evidence that the rocks were subjected to post-crystalline deformation.

Chess-board albite has also been described from greywackes (Cummins, 1957) but whether this is detrital chess-board albite or detrital albite (possibly secondary after potassic feldspar ?) in which the chess-board pattern has developed subsequent to deposition is not clear.

Acknowledgments

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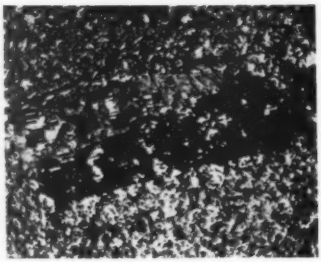
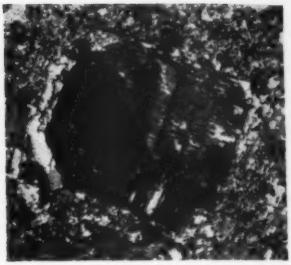
FIG. 1.



3·0 mm.

FIG. 3.



1·0 mm.

3·0 mm.

FIG. 2.

1·0 mm.

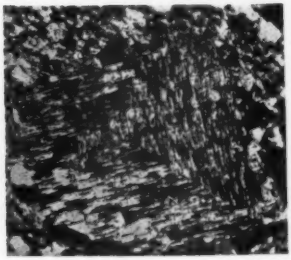


FIG. 4.

CHESS-BOARD ALBITE

useful advice and criticism; also to Dr. D. Flinn for his help in the early stages of the work. I would also like to express my sincere thanks to Mr. J. Stringer and Mr. J. Preston of the Department of Metallurgy, The University of Liverpool, for their valuable assistance with the photography.

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THE JANE HERDMAN LABORATORIES OF GEOLOGY,
THE UNIVERSITY OF LIVERPOOL.

DESCRIPTION OF PLATE III

CHESS-BOARD ALBITE

FIG. 1.—A photomicrograph of perthite showing Carlsbad twinning.

FIG. 2.—A photomicrograph of chess-board albite showing Carlsbad twinning, this is difficult to discern due to the complexity of the chess-board twinning but the lower half of the crystal is slightly darker.

FIG. 3.—A photomicrograph of perthite showing Baveno twinning.

FIG. 4.—A photomicrograph of chess-board albite showing Baveno twinning.

Six New Genera of Mesozoic Brachiopoda

By Graham F. Elliott

Abstract

Six new genera of terebratellacean brachiopods are proposed; one Triassic, one Jurassic, and four Cretaceous.

In connection with current work on Mesozoic brachiopods of the superfamily Terebratellacea Allan (1940), for the *Treatise on Invertebrate Paleontology*, six new genera are now proposed. Two of these are foreign, and four are familiar British Cretaceous fossils.

Strictly speaking, the attribution of a genus to the terebratellaceans should be based on a knowledge of the fact that its loop develops in connection with a median septum, and a much fuller knowledge of the development of certain other long-looped forms is also desirable (Elliott, 1953, Muir-Wood, 1955). Such knowledge depends on the preservation and discovery of exceptional material suitably preserved for developmental study; in the meantime, some attempt must be made to classify the many fossil forms for which such direct evidence is lacking. The present six are foreign forms of special interest, or English forms well-known to the writer; they are but a small sample of numerous described species of Cretaceous *Terebratella*, etc., which will eventually have to receive similar treatment.

Family Dallinidae Allan 1940

1. *Eodallina* gen. nov.; type-species *E. peruviana* sp. nov. ("terebratellacean species" of Stehli (1956), p. 102, figs. 1–8). Shell small, smooth, rounded-ovate, biconvex, with shallow brachial sulcus only; beak straight, entire, pedicle-opening triangular without deltidial plates. Pedicle-valve interior with short strong dental lamellae, and a medianly divided ventral muscle field. Brachial interior with strong sockets and inner socket-ridges enclosing concave hinge-trough supported by low median septum; this extends anteriorly past mid-length, supporting a spinous modified campagiform loop with short crural points, broadly-attached descending branches and a hood thinning to a narrow transverse band. Upper Triassic; Peru, South America: the earliest terebratellacean known.

2. *Antigoniarcula* gen. nov.; type-species *Ismenia perrieri* (E. Eudes-Deslongchamps, 1884) = *Argiope perrieri* E. E.-Desl. 1853. Shell small, transverse, alate, test ornamented by numerous more or less equal costae, new ones being intercalated with increase of size, and with close-set scaly growth-lines crossing costae and intercostae alike; a shallow median sinus in the dorsal valve opposed by an opposite fold in the ventral valve. Area low, wide, set back, foramen large

with narrow deltidial plates and pedicle-collar; lateral areas wide, with transverse growth-lines; hinge-line megathyrid. Brachial interior with sub-central sockets, a small hinge-plate delimited by inner socket-ridges and supported anteriorly by a very short thin median septum, and showing a small cardinal process. Loop angular dalliniform, anteriorly produced into sharp points, and with thin and delicate crura. Middle Lias, Lower Jurassic; France and England.

3. *Oblongarcula* gen. nov.; type-species *Terebratula oblonga* J. de C. Sowerby, 1826. Shell medium-sized, elongate-ovoid, biconvex, commissure rectimarginate to sulcate, test thin, ornamented by fine regular costae bifurcating only near the unbo. Beak sub-erect, foramen mesothyrid, deltidial plates conjunct, lateral areas delimited by beak-ridges, hinge-line terebratulid. Pedicle-valve interior with strong hinge-teeth, and thin dental lamellae close to the sides of the valve. Brachial-valve interior with thin plate-like cardinalia, cardinal process raised transverse, sockets narrow, inner socket-ridges enclosing wide thin hinge-plate supported below by median septum. Median septum thin, plate-like, extending anteriorly to half the valve-length, crura delicate, loop believed to be terebrataliform. Lower Cretaceous; Europe.

4. *Arenaciarcula* gen. nov.; type-species *Terebratella fittoni* Meyer 1864*b*. Shell similar to *Oblongarcula* but smaller, with fewer and coarser costae, thicker test, commissure uniplicate to parasulcate, dental lamellae fused to sides of valve, and cardinal platform with well-marked raised separate inner socket-ridges, crural bases and median septal ridge. Loop believed terebrataliform. Lower Cretaceous; Western Europe.

5. *Trifidarcula* gen. nov.; type-species *Terebratella trifida* Meyer 1864*a*, *b*. Shell small, transverse, test thick with three principal high raised rounded dorsal folds alternating with two ventral folds; foramen large, open, deltidial plates small, area sloping, hinge-line megathyrid. Pedicle-valve interior with thick teeth with supports fused to valve. Brachial-valve interior with small thick cardinal platform of fused elements, cardinal process small; medium septum thick, extending anteriorly for half the valve-length; loop believed to be terebrataliform. Lower Greensand, Aptian, Lower Cretaceous; England.

Family Terebratellidae King 1850

6. *Dereta* gen. nov.; type-species *Terebratula pectita* J. Sowerby 1816. Shell medium to large, transverse terebratuliform, strongly biconvex, test thick, ornamented by numerous bifurcating striae, commissure sulcate. Area wide and high, foramen round to oval, encroaching slightly on gently incurved beak, lateral areas smooth, on either side of symphytium, beak-ridges sharp, hinge-line subtere-

bratulid. Pedicle-valve interior with large thick teeth, without dental lamellae, but pedicle-collar present and median ridge in umbonal area. Brachial-valve interior with small thick cardinal platform of fused elements, dominated by a high narrow pillar-like cardinal process; septum thick, loop believed terebratelliform. Middle Cretaceous; Europe.

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IRAQ PETROLEUM CO.,
214 OXFORD STREET,
LONDON, W. 1.

Mechanics of Jointing in Rocks

By Neville J. Price

Abstract

The current theories dealing with the mechanics of joint development are discussed critically. An analysis of the stress conditions which can exist in competent rocks in the crust is presented. The conclusion is reached that " Master " and " Regional " Joints can be, generically, either " tensional " or " shear " structures. They are, however, post-tectonic phenomena which develop as a result of uplift. Joints form a part of " movement pictures " because their orientation is determined by residual stresses. Reasons are given which indicate that the mechanism described is generally applicable to joint development in folded sediments, igneous rocks, and also to incompetent material.

I. Introduction

JOINTS in rocks have long been studied; their characteristics and mode of occurrence have often been described and several theories regarding the origin of joints, particularly " diastrophic joints " (i.e. those due to tectonic processes) have been proposed. However, much doubt still exists regarding the actual mechanics of joint formation.

Diastrophic joints are defined as fracture planes, along which there has been little or no displacement. The fractures are usually planar and the joint surfaces are relatively smooth. In sediments which are only slightly deformed the joint planes are often vertical and are generally uniform in direction over large areas; these structures can be termed " Regional Joints ". Two complementary sets frequently develop and three, or even four, distinct sets are not uncommon. When deformation has been considerable the joints often form an important part of the " movement picture ". The relationship of the main joint directions to the movement picture has been described by Wilson (1952) and de Sitter (1956). They point out that ideally the joints occur in two systems. The first comprises the " ac " or " Cross Joints " and the complementary set of " Longitudinal Joints " parallel to the " b " direction. These joints are frequently classified generically as " Tension Joints ". The second system comprises the two sets of complementary " Oblique Joints ", often termed " Shear Joints ", which make an angle of less than 45° with the " a " axis. It should be noted, however, that slight variations in the direction of these sets and the introduction of sets of joints unrelated to the " movement picture " can result in confusion.

One of the earliest theories dealing with joint development suggests that these structures are " tensile " fractures. The lack of movement parallel to the joint plane is evidence in favour of this theory. Moreover, there is little doubt that some radial joints at the crest of a fold

developed as a result of tensile stresses which can occur during folding. It is difficult, however, to envisage " Oblique Joints " as being the direct result of tensile stresses acting normal to the joint planes.

Following their study of structures in the South Limburg Coalfield, Van Waterschoot Van de Gracht *et al.* (1942), concluded that joints in rocks and cleats in coal were shear structures which formed as the result of horizontal compression and were among the earliest structures to develop. One of the criticisms which can be levelled against this theory was put forward by Kendall and Briggs (1933). They pointed out that while the compression and shear theory explains the orientation and development of faults it does not offer any satisfactory reason for the large number of joints developed in rocks or the lack of displacement along individual joint planes.

Instead of compression and shear, Kendall and Briggs advocated a torsional theory of joint formation. They considered that joints were fatigue phenomena which developed early in the history of the sediments and resulted from alternating, regularly orientated, torsional stresses brought into play by semi-diurnal tidal action. This theory, however, is completely unable to explain the relationship which exists between joints and the " movement picture " of folded sediments.

A criticism which can be applied to both the shear and the torsional theories is based upon the observation that movement along joint planes and deformation of the plane is rarely observed. Thus, in general, it seems unlikely that joints formed early in the tectonic history of sediments. Turner (1948) has further argued that no rock undergoing plastic deformation can at the same time develop joints, for rock flow is apt to eradicate any jointing the rock may possess. Consequently, it seems certain that, in metamorphic rocks at least, joints formed after the main tectonic compression when the rocks had become less plastic. From this it can be inferred that joints developed at higher levels in the crust than those at which the rocks underwent tectonic deformation.

In the paragraphs which follow, a theory is presented which resolves the difficulties noted above. The stress conditions which can obtain in competent rocks during various periods of their tectonic and post-tectonic histories are considered. It is then shown that, in general, neither " Tension " nor " Shear " Joints can develop during a compressive phase and that jointing occurs because of the lateral expansion and tensile stresses which develop as a result of uplift. The argument is initially restricted to " Regional " and " Master " Joints, but reasons are later presented which indicate that the mechanism suggested can be applied to jointing in igneous rocks and to relatively small scale jointing in folded sediments. Joint development in incompetent material is also discussed.

II. Gravitational and Tectonic Stresses in Competent Rocks

In this simple treatment it is assumed that competent rocks are brittle materials which are completely elastic up to the point of failure. This assumption is supported by some laboratory experiments, such as those of Handin and Hager (1957), in which it is shown that competent rocks may fail as brittle material at loads as high as 140,000 lb. in.$^{-2}$ when acted upon by a confining pressure of 28,000 lb. in.$^{-2}$ An objection to treating rocks as simple elastic bodies may be raised on the grounds that, in general, rocks exhibit evidence of permanent deformation, the result of plastic, or viscous flow. Now Scheidegger (1957) considers that the persistence of mountain-chains is evidence that rocks possess a minimum yield strength which must be exceeded before flow takes place; i.e. he believes rocks to approximate to "Bingham Bodies". But, as de Sitter (1956) has pointed out, tectonic processes are slow and in general the stresses which develop only slightly exceed the yield strength, or in other words, for Bingham Bodies, the stresses are only slightly greater than those which result from elastic strain. Consequently, the following treatment based on elastic theory represents quite closely the stress conditions which can obtain in competent rocks. It is pointed out that since the brittle behaviour of materials is dependent upon temperature as well as pressure, the assumption made will only hold at the relatively low temperatures obtaining in the upper 20,000–30,000 feet of the crust.

The basic elastic equations relating triaxial stress and strain are:—

$$e_x = \frac{1}{E}\left[\sigma_x - \frac{1}{m}(\sigma_y + \sigma_z)\right] \quad . \quad . \quad . \quad (1)$$

$$e_y = \frac{1}{E}\left[\sigma_y - \frac{1}{m}(\sigma_x + \sigma_z)\right] \quad . \quad . \quad . \quad (2)$$

$$e_z = \frac{1}{E}\left[\sigma_z - \frac{1}{m}(\sigma_x + \sigma_y)\right] \quad . \quad . \quad . \quad (3)$$

where $e_{x,y,z}$ are the strains in mutually perpendicular directions,

$\sigma_{x,y,z}$ are the corresponding stresses,

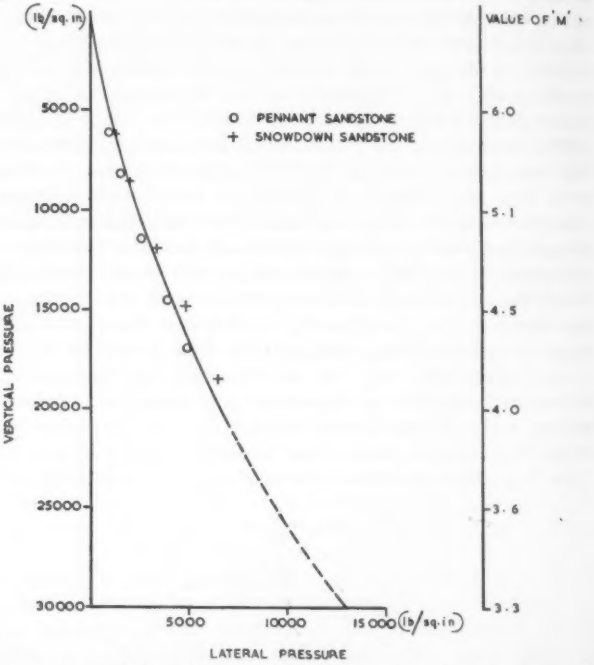
E is Young's Modulus, and

m is Poisson's Number, the reciprocal of Poisson's Ratio.

These equations can be used to establish the stress conditions in the earth's crust. Thus, consider a small unit cube of rock in the crust subjected to a vertical gravitational stress σ_z. Lateral expansion is completely restricted by the pressure of the surrounding rock; i.e. $e_y = e_x = 0$ and, therefore, from equations (1), (2), and (3) it follows that the lateral pressure is given by:—

$$\sigma_y = \sigma_x = \frac{1}{m-1}\cdot\sigma_z \quad . \quad . \quad . \quad . \quad (4)$$

The value of m varies from rock type to rock type. Phillips (1948) has shown that the value of m is also a function of pressure. He showed, in uniaxial experiments, that the value of m decreases as loading increases. Price (1958) has established a similar relationship in con-



TEXT-FIG. 1.—Relationship between vertical and horizontal pressure in typical competent sandstones also showing values of Poissons Number at the various stress levels.

ditions of triaxial compressive stress. The values of m and the manner in which horizontal pressure is related to the depth of burial for two moderately competent coalmeasure sandstones, each with a uniaxial crushing strength in excess of 20,000 lb. in.$^{-2}$, are shown in Text-fig. 1.

Consider now the effect of a horizontal stress c_y superimposed upon the stress field due to gravity. Then, neglecting the effect of the very small increase of overburden which results from the compressive stress c_y, the new stresses in the y and x direction are given by:—

$$\sigma'_y = \frac{1}{m-1} \cdot \sigma_z + c_y \quad . \quad . \quad . \quad . \quad (5)$$

and $$\sigma'_z = \frac{1}{m-1} \cdot \sigma_z + \frac{1}{m} \cdot c_v \quad . \quad . \quad . \quad (6)$$

Clearly σ_z, σ'_y, and σ'_z are principal stresses and when their relative values are known the greatest, intermediate, and least principal stresses can be designated by σ_1, σ_2, and σ_3 respectively.

It will be seen that all the principal stresses are compressive, so it can be concluded that Regional and Master Tension Joints cannot develop during such a compression.

However, as previously noted, joints exhibit some of the characteristics of shear fractures and are possibly related to wrench or transcurrent faults. It is of interest, therefore, to establish the stress conditions which must exist before such structures can develop.

Relatively homogeneous, brittle rocks apparently fail in accord with Mohr's Theory of Fracture and, therefore, two conditions must be satisfied before wrench shear planes can develop:—

(1) The intermediate principal stress must act vertically at the instant of failure.

(2) The shear stress, i.e. $(\sigma_1 - \sigma_3)/2$, must be sufficiently large to overcome the resistance to shearing.

From equations (5) and (6) it follows that if the load due to gravity is to represent the intermediate principal stress, i.e. $\sigma_z = \sigma_2$, then the value of the compressive stress c_v must fall within the limits represented by:—

$$\sigma_z \cdot \frac{(m-2)}{(m-1)} \leqslant c_v \leqslant \sigma_z \cdot \frac{m.(m-2)}{(m-1)} \quad . \quad . \quad . \quad (7)$$

Obviously the limiting conditions of c_v are determined by the value of m. Thus, if $m = 4$, in order that the intermediate principal stress may act vertically c_v must fall between the limiting values $0{\cdot}66.\sigma_z$ and $2{\cdot}66.\sigma_z$. By substituting these values in equations (5) and (6) it can be shown that the ratio of the greatest to least principal stresses (σ_1/σ_3) cannot exceed 3·0. The maximum ratio of σ_1/σ_3 for other representative values of m are given in Table 1 (p. 154).

It is now necessary to establish the conditions which must obtain before shear fractures will develop and to see whether they are compatible with the one just established. According to Coulomb's Equation, upon which Mohr's Theory is based, shear fractures develop when:—

$$\tau = C + \mu\sigma_n \quad . \quad . \quad . \quad . \quad (8)$$

where τ is the shearing stress,

σ_n is stress normal to the shear plane,

C is the cohesive strength of the material,

and μ is the coefficient of internal friction.

While the angle at which shears develop, relative to the axis of greatest principal stress, is given by:—

$$\alpha = \mp 45° \pm \omega/2 \quad . \quad . \quad . \quad . \quad (9)$$

where

$$\omega = \mathrm{Tan}^{-1}\mu \quad . \quad . \quad . \quad . \quad . \quad (10)$$

In this analysis it is assumed that the value of μ does not change with pressure. [Laboratory experiments have in fact shown that for many competent rocks the coefficient of internal friction does not alter appreciably with pressure.] The value of this quantity has been inferred from " field evidence " (see Anderson, 1951) as well as being determined experimentally; in both instances the values range from about 0·84 to 1·38 for competent rocks.

TABLE 1.

The maximum ratio of greatest and least principal stresses which can develop if the intermediate principal stress is to act vertically for various values of m.

Value of Poisson's Number m	Maximum Ratio σ_1/σ_3
2	1·0
3	2·0
4	3·0
5	4·0
6	5·0
7	6·0

Now the normal and shearing stresses which develop on any plane are given by:—

$$\sigma_n = \sigma_1 . \mathrm{Sin}^2\phi + \sigma_3 . \mathrm{Cos}^2\phi \quad . \quad . \quad . \quad (11)$$

and

$$\tau = (\sigma_1 - \sigma_3) . \mathrm{Sin}\phi . \mathrm{Cos}\phi \quad . \quad . \quad . \quad (12)$$

where ϕ is the angle any plane, orientated parallel to the axis of intermediate principal stress, makes with the axis of greatest principal stress. By using equations (8), (9), (10), (11), and (12) it is possible to establish the relationship which must exist between the least and greatest principal stresses before shearing can take place.

If it is assumed that the cohesive strength of the material is negligibly small—as may be the case in prejointed rocks—the ratio of σ_1/σ_3 can be readily determined. The ratios of the greatest and least principal stresses necessary to induce failure in rocks with zero cohesive strength and coefficients of internal friction ranging from 0·84 to 1·38 are given in Table 2. The cohesive strength of competent rocks, however, is usually of the order of 4,000–5,000 lb. in.$^{-2}$ (280–350 kg. cm.$^{-2}$) and will greatly influence the conditions which must be satisfied before failure

will occur. To illustrate the influence of this factor it is necessary to give specific values either to the greatest or the least principal stress. The ratios of σ_1/σ_3 necessary to bring about failure in a rock with a coefficient of internal friction of 0·84 and a cohesive strength of 4,000 lb. in.^{2} are given in Table 3. These figures will give approximately the lowest ratios of σ_1/σ_3 necessary for shear failure in competent rocks.

TABLE 2.

Ratios of greatest to least principal stresses necessary to cause shearing in rocks with zero cohesive strength for different values of internal friction.

μ	σ_1/σ_3	Angle of Fracture
1·38	35·0	36°
1·00	6·0	45°
0·84	4·5	50°

TABLE 3.

Ratios of greatest and least principal stresses necessary to cause shearing in a rock with a coefficient of internal friction of 0·84 and a cohesive strength of 4,000 lb. in.^{2}

σ_1	σ_3	σ_1/σ_3
17,000 lb. in.^{2}	0 lb. in.^{2}	∞
63,000 „ „	10,000 „ „	6·3
110,000 „ „	20,000 „ „	5·5

Bearing in mind the point that tectonic compression decreases the value of *m*, then if the data in Tables 1, 2, and 3 are compared it will be seen that the conditions which must be satisfied before shearing can occur are not compatible with those which must obtain if the intermediate principal stress is to act vertically. Consequently it can be concluded that wrench shear fractures will not develop in the conditions postulated. One is forced to the interesting conclusion that such shears can only develop if there is a lateral expansion, concomitant with a lateral compression, which has the effect of increasing the ratio of the greatest to least principal stresses to such an extent that shearing can occur. Wrench faults are somewhat rarely observed, which, in the light of the previous conclusion, is understandable. Joints, however, are ubiquitous and it must be concluded, therefore, that Regional and Master Oblique Joints did not form during the main phase of tectonic compression. It is necessary, therefore, to follow the investigation further and consider the possible sequence of events and the stresses

which act upon competent rocks during their post-deformational histories.

III. Post-deformational History

During the deformational phase the resistance to compression of rocks in the crust builds up until it is in equilibrium with the maximum available compressive forces; further strain ceases when these conditions are attained. If the rocks are ideal Bingham Bodies the elastic strain energy remains stored in the rock. Thus, while there is no further movement, " residual stresses " are in existence in the rocks which faithfully represent in quantity and direction the stresses which acted at the end of the tectonic phase. If, however, rocks can be considered as " Elastico-Viscous Bodies ", then over a long period of time the stress differentials between the various principal stresses will disappear: the residual stresses will " relax " resulting in a state approaching " Hydrostatic Stress ".

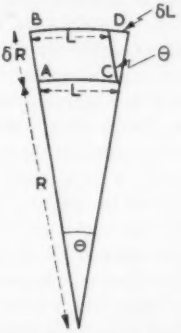
Data regarding " relaxation time " (i.e. the time taken for residual stresses to relax to $1/e$ of their original magnitude) are exceedingly scarce. Scheidegger (1957) calculated that " the crust has a relaxation time of 20,000 years upwards ". This figure is based upon the value of Maxwell Viscosity obtained by Gutenberg (1941) from data referring to the post-glacial uplift in Fenno-Scandia. The data clearly refer to an " average " rock type situated deep within the crust at elevated temperatures. It is virtually certain that the viscosity of competent rocks in the cooler upper layers of the crust will be considerably higher.

Griggs (1939) carried out measurements of " creep " in Solenhofen limestone in order to establish a value for the viscosity of a competent rock. Viscous flow results in a non-linear " Strain–Log time " relationship. In his experiments Griggs detected no systematic deviation from a straight line during the 550 days of test. The calculated coefficient of viscosity of $2{\cdot}2 \times 10^{22}$ poises, which agrees well with the figure obtained by Gutenberg, is based upon the maximum spread of his observations. As Griggs himself has pointed out the figure quoted is the minimum possible value of viscosity because no correction has been applied for " creep " in the apparatus. Furthermore, the spread of results include experimental errors such as may be incurred by small fluctuations in temperature and humidity. Thus, it is by no means certain that Griggs has demonstrated that the limestone exhibited viscous flow. But, even if the rock possesses a viscosity under the test conditions, the coefficient will almost certainly be in excess of 10^{23} poises. It follows then that the relaxation time of such a rock, in these conditions, is not less than 100,000 years and could quite easily be greater than 10,000,000 years. If this latter figure is taken as a reasonable estimate of the relaxation time it would take a period of

some 50,000,000 years for residual stresses to relax to 1 per cent of of their original value.

In subsequent paragraphs the cycle of events in the post-deformational history attendant upon uplift and erosion of the superincumbent strata is considered in relation to the two possible end conditions, i.e. (1) hydrostatic pressure and (2) unrelaxed tectonic stresses.

Joints are observed in rocks which are encountered at the surface or in mines at relatively shallow depths. All the rocks thus encountered have at one time or other been buried under later sediments, compacted and compressed and, possibly after a period of tectonic compression, lifted up and the superincumbent material removed by erosion



TEXT-FIG. 2.—Simple geometrical illustration of " Plane Strain " extension which takes place if a line (representing a bed) is uplifted from *CA* to *BD*.

and denudation. This decrease in cover is attended by a number of effects which include (1) a decrease in gravitational loading, (2) a proportional decrease in the lateral stresses (See Text-fig. 1 and Eq. 4–6), and (3) a lateral strain which develops in the rocks.

Consider this latter point. In general a unit of competent rock will have considerable lateral extent and can be represented in section as an arc of radius R from the centre of the earth. If uplift takes place the radius becomes $(R + dR)$. But (see Text-fig. 2):—

$$L = R.\theta \quad . \quad . \quad . \quad . \quad (13)$$

and, therefore, there is a lateral strain, which is the result of uplift, given by:—

$$dL/L = dR/R \quad . \quad . \quad . \quad . \quad (14)$$

Assuming the value of dR is approximately 20,000 feet, or 4 miles, then, since $R = 4{,}000$ miles

$$dR/R = dL/L = 1/1{,}000$$

This effect is represented in Text-fig. 2 as " plane strain ". The strain is in fact, of course, radial in the horizontal plane.

If this lateral expansion occurs by elastic stretching, a tensile stress (T) tends to develop, of value $E.dL/L$ (where E is Young's Modulus). The author has shown (1958) that the value of Young's Modulus is dependent upon pressure, but for the coalmeasure rock types previously mentioned the modulus parallel to the bedding is in excess of 1×10^7 lb. in.$^{-2}$. Using these values, the tensile stresses which tend to develop in these rocks as a result of an uplift of 20,000 feet is 10,000 lb. in.$^{-2}$. Now gravitational loading increases by approximately 1 lb. in.$^{-2}$ for every foot of increase of depth of cover. Thus, the vertical load acting upon rock which has been lifted up 20,000 feet to the surface has decreased from 20,000 lb. in.$^{-2}$ to zero; therefore, the tensile stresses which tend to develop in this period of uplift are equal to half the change in gravitational load. For ease of presentation it is henceforward assumed, in this paper, that this approximate relationship is general to all competent rocks at the depths being considered.

One is now in a position to consider the variations in the conditions of stress which occur in a rock as a result of uplift. Two examples will be considered: (*a*) it will be assumed that relaxation of stresses has taken place and that the rocks are subjected to a hydrostatic, or near hydrostatic compression, while in the second example (*b*) it will be assumed that the tectonic stresses are not relaxed. It will be shown that the tensile stresses which tend to develop during uplift are vitally important and can give rise to the stress conditions necessary for the development of " Shear " and/or " Tension " Joints.

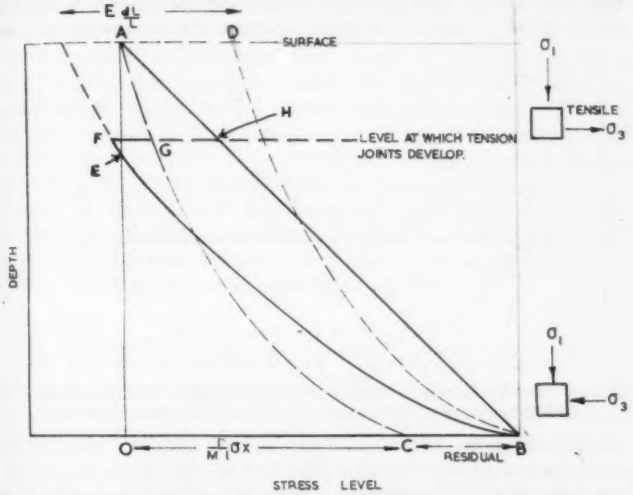
(*a*) *Initial Conditions Approximate to the Hydrostatic State.*

It is here assumed that initially the stress conditions obtaining in the crust are hydrostatic and that uplift and erosion of the superincumbent material takes place at such a rate that subsequent stress differences are largely unaffected by further relaxation.

The initial and subsequent stress conditions are represented graphically in Text-fig. 3. The straight line *AB* represents the gravitational loading which, of course, is directly proportional to depth. The lateral stresses are represented by the gravitational and residual components *OC* and *BC*. As uplift takes place, the lateral component of stress due to gravity decreases in the manner similar to that indicated by *CGA* (cf. Text-fig. 1) so that the decrease in the total lateral stress due to this factor is represented by *BD*. As noted earlier, it is assumed that the tensile stresses $\left[T = \frac{EdL}{L}\right]$ are equal in value, but opposite in sign, to half the decrease in vertical load which is brought

about by uplift. Thus, the resultant variations in the lateral stresses which take place in competent rocks is represented by *BEF*.

It will be seen that the total lateral stresses rapidly decrease so that the gravitational load represents the greatest principal stress ($\sigma_z = \sigma_1$). These are the conditions conducive to the formation of normal faults; and indeed such faults may develop in incompetent rocks associated with the competent material under consideration. It must be noted, however, that the maximum depth represented by Text-fig. 3 is about



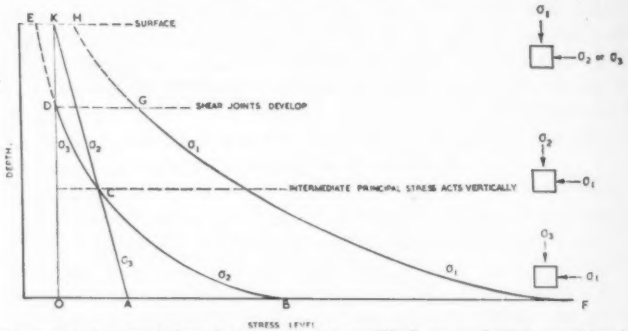
TEXT-FIG. 3.—Variations in stress due to uplift from an initial condition of near hydrostatic stress.

30,000 feet and the greatest pressure, the initial condition, approximately 30,000 lb. in.$^{-2}$. The maximum stress difference ($\sigma_1 - \sigma_3$) is represented by *FH* and is approximately equal to 10,000 lb. in.$^{-2}$ which is too small to cause shearing to develop in competent rocks at this depth in the crust.

At point *E* the lateral stresses are zero and at shallower depths the rocks actually go into tension. The Tensile Strength of competent rocks is usually of the order of 1,000–3,000 lb. in.$^{-2}$ so that further uplift will result in the formation of vertical tension joints. Structures thus formed will have a random orientation and a joint pattern similar to that exhibited by " columnar basalt " may well result.

In general, however, it is to be expected that the initial stress conditions will only approximate to the hydrostatic state. The lateral stresses will not necessarily be equal so that, on uplift, one will represent

the least and the other the intermediate principal stress. King-Hubert and Willis (1957) have shown experimentally that tension joints, when they develop, form at right angles to the axis of least principal stress, so that when the stress conditions indicated by point *F* (Text-fig. 3) are attained joints with a definite orientation will develop. The moment fracture takes place the least principal stress changes, the tensile stresses are released and are replaced by a compressive stress with a value $\sigma_z . 1/(m-1)$, cf. point *G*. The former intermediate principal stress thus becomes the least principal stress. Further uplift will cause a second set of " Tension " Joints to develop; the two sets forming an



TEXT-FIG. 4.—Variations in stress due to uplift from an initial condition in which the ratio of greatest to least principal stresses (σ_1/σ_3) is large.

orthogonal system. If the lateral stresses were initially almost exactly equal, the formation of the sets of joints will be practically simultaneous. The orientation of the residual stresses, even after a long period of relaxation, will reflect the original stress trajectories which existed during tectonic compression. Since these tensions form normal to the horizontal axes of principal stress, they will be " ac " or " Cross " Joints and " Longitudinal " Joints.

(*b*) *Initial Condition in which Tectonic Stresses are not Relaxed.*

The lateral stress conditions which arise during uplift, if the original tectonic stresses do not relax, are given by:—

$$\sigma_y = \frac{1}{m-1} . \sigma_z + c_y - T \quad . \quad . \quad (15)$$

and

$$\sigma_x = \frac{1}{m-1} . \sigma_z + \frac{1}{m} . c_y - T \quad . \quad . \quad (16)$$

where $T = \frac{EdL}{L}$, the tensile stress which is the result of the uplift.

The initial and subsequent stress conditions are represented in

Text-fig. 4 which, it should be noted, has, relative to Text-fig. 3, a four times vertical exaggeration.

The initial condition represents the stresses which are residual after a period of thrusting, the greatest principal stress which, at this stage, acts horizontally. As uplift takes place the greatest and intermediate principal stresses change in the manner indicated by curves *FG* and *BCD*. The reasons for these changes have been discussed in the previous example. At the level indicated by *C* the vertical load changes from the least principal stress to the intermediate principal stress, so that one set of conditions conducive to the formation of transcurrent or wrench fractures is satisfied. It can be inferred from Text-fig. 4 that as uplift progresses further the ratio of the greatest to least principal stresses increases rapidly until the second set of conditions necessary for the development of shear fractures are fulfilled at or near the level indicated by *DG*. Immediately, shear joints develop and the residual stresses are largely dissipated so that gravitational loading assumes the role of greatest principal stress. These conditions are, of course, similar to those discussed in Example 1. Further uplift will cause the rocks to pass into tension in the horizontal direction so that a system of Tension Joints may subsequently develop.

IV. Sets of Joints

It has been shown that two sets of " Tension " Joints will develop in rocks which have undergone only slight tectonic compression, or if the tectonic stresses have relaxed. Similarly, two sets of Master " Shear " Joints and two sets of Master " Tension " Joints can develop in rocks which have undergone considerable tectonic compression. It is emphasized that these structures develop as a result of one cycle of subsidence, compression, and uplift. When a larger number of sets of joints have developed than the number specified above, it is reasonable to conclude that the rocks have undergone more than one cycle of subsidence and uplift.

Prior to, or accompanying, uplift it is possible that the simple sequence of events so far considered may be complicated by regional lateral expansion or compression which may be in a direction unrelated to the earlier compression. Such extension or compression will then upset the existing stress conditions. For example, extension in the general direction of the least principal stress will facilitate the formation of " Shear " Joints; but extension in the direction of the greatest principal stress will impede, or even prevent, the development of such joints, so that only " Tension " Joints form. Furthermore, a complete reorientation of the residual stresses is probable so that joints which subsequently develop will be completely unrelated to the original " movement picture ".

V. Depth at which Joints Develop

It has already been pointed out that the maximum depth represented by Text-figs. 3 and 4 is about 30,000 feet. From the position of the level *FG*, in Text-fig. 3, it can be concluded that " Tension Joints " form in a rock which has undergone only moderate tectonic compression at a depth of some 8,000–10,000 feet below the surface. " Shear Joints " in highly compressed sediments develop at a similar depth below the surface. The subsequent development of " Tension Joints " in such rocks must occur at a shallower depth.

Since sediments are deposited below sea-level and are usually observed, in the field, at some distance, perhaps many thousands of feet, above sea-level rather more uplift is likely to occur than has been assumed in this paper. Consequently, joints may form at a greater depth, relative to the surface, than is indicated above.

VI. Joint Frequency

Anderson (1951) has shown that movement along a fault plane has the effect of shortening a faulted block in the direction of the greatest principal stress and increases the extent of the block in the direction of the least principal stress. The shearing movement along the plane may be considerable and stress conditions initiating this movement can be relieved over a wide area.

The state of affairs which obtain prior to the formation of " Shear Joints " is different; the stresses in the rock are residual and cannot be replenished by tectonic processes. The stress deviator has a value of approximately 20,000–30,000 lb. in.$^{-2}$ which can be completely dissipated by a movement of a few thousandths of an inch along a shear plane. Such a movement would, however, only dissipate the residual stresses in the immediate vicinity of the joint plane. Consequently, a regional residual stress can only be dissipated by the formation of a very large number of joints. A similar argument applies to the number of tension joints which develop as a result of uplift.

Not all rocks exhibit the same joint frequency. For example, joints (i.e. cleats) in coal occur very much more frequently than joints in a nearby sandstone (several cleats to the inch as opposed to one joint every foot or so in the sandstone). The author believes that the number of joints which develop in a rock is related to the strain energy originally stored in the rock. The strain energy stored in an elastic body is equal to the work done in producing a given amount of strain. Thus, in a unit cube of material which has a linear stress-strain characteristic:

$$w = \tfrac{1}{2}.\sigma.e \quad . \quad . \quad . \quad . \quad (17)$$

where w is the strain energy
σ is the applied stress
and e is the resulting strain.

But, $e = \sigma/E$ (where E is Young's Modulus), therefore:—

$$w = \tfrac{1}{2}.\sigma^2/E \quad . \quad . \quad . \quad . \quad (18)$$

From equations (1), (2), (3), and (18) it can be shown that the strain energy stored in a unit cube subjected to triaxial compression is given by:—

$$W = \frac{1}{2E}\left[\sigma_1^2 + \sigma_2^2 + \sigma_3^2 - \frac{2}{m}.(\sigma_1\sigma_2 + \sigma_1\sigma_3 + \sigma_2\sigma_3)\right] \quad . \quad (19)$$

From this last equation an estimate can be made of the strain energy stored in various rock types. For example, if coal (with $E = 2 \times 10^5$ lb. in.^{2} and mean value of $m = 3$) and sandstone (with $E = 1 \times 10^7$ lb. in.^{2} and mean value of $m = 5$) are both subjected to a vertical load of 10,000 lb. in.^{2} and a compressive stress (σ'_v) of 20,000 lb. in.^{2} then the strain energy stored in the coal is 32·5 times greater than that stored in the sandstone. Thus, the ratio of joint frequency and strain energy are of the same order. It is emphasized that equation (19) only gives an estimate of the strain energy, for the stress-strain relationships of rocks is rarely linear. Moreover, it is not known at what stage of compaction and compression that the rocks attained their ultimate elastic properties.

VII. Jointing in Folded Sediments

It has so far been assumed, for the sake of simplicity, that the principal stresses act vertically and horizontally. King-Hubbert (1951) has shown, however, that in folded belts the stress trajectories are not in general horizontal and vertical. The orientation of the stress trajectories during the tectonic phase will influence the orientation of the joint planes which subsequently develop, hence, the linear intersection of Master Joints need not be vertical in folded rocks.

An analysis of the stresses induced locally by the mechanics of folding is beyond the scope of this paper. However, as stated earlier, radial "tension" joints, which are often quartz filled, can develop near the crest of a fold during flexure. In interbedded competent and incompetent beds, tensile fractures can also form in the harder material (cf. the conditions which give rise to boudinage). In general, however, in the limbs of a fold, away from the crest, the conditions previously established must be satisfied before "Shear" and "ac" Tension Joints can develop. Therefore, as with Master Joints, the majority of minor joints will develop during the subsequent period of uplift. The inclination of stress trajectories within a fold is largely controlled by the orientation of the competent beds, so that minor joints will, in general, form perpendicular to the bedding.

VIII. Jointing in Igneous Rocks

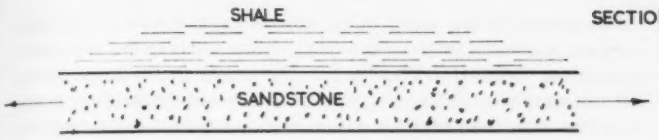
Joints can only form in igneous rocks after the solid state has been attained. The presence of quartz and other material in some joint planes indicate that these structures formed at a relatively early period in the " solidification " of the material and that such joints may best be attributed directly to the mode of emplacement. However, the mode of formation of " Q " and " S " Joints (i.e. Cross and Longitudinal Joints with the typical orientation described by Balk), is not clearly understood. It is suggested here that the formation of these joints is satisfactorily explained by the mechanism described in previous paragraphs. They are, therefore, " Tension " Joints which developed in the solidified mass as a result of the lateral stretching attendant upon uplift, the orientation of the structures being determined by the stress conditions which were in existence during emplacement of the rock.

IX. Jointing in Incompetent Material

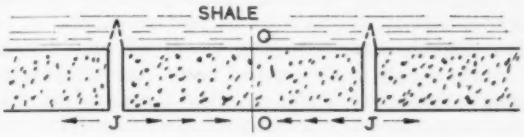
Joints are often poorly developed in thick, incompetent members. This can probably be related to the fact that the elastic moduli of incompetent material are smaller in value than those for competent rocks. As a result, the tensile stresses which tend to develop during uplift are comparatively small. Also, the existing evidence indicates that shaly material may be " Elastico-Viscous ", with a fairly small viscosity, so that stress differentials can be expected to relax in a relatively short period of time. The occurrence of irregular, curved fracture planes which are sometimes observed in such material can probably be attributed to the absence of a definite pattern of residual stresses.

The joints which occur in the weaker material of a series of interbedded competent and incompetent rocks can be attributed to the influence of the stronger material. This, it is thought, can be brought about in one of two ways.

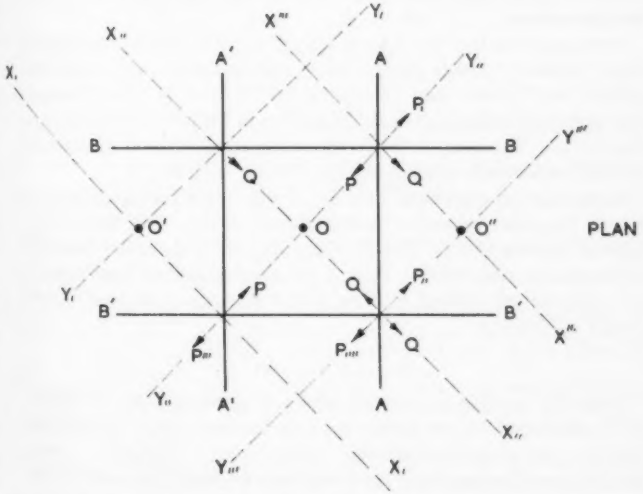
Firstly, Inglis (1913) has shown that very large tensile stresses can occur at the end of a fracture so that a fracture, or joint, once formed is largely self-propagating. Thus, in a series of rocks in which competent grits grade up into incompetent shales or mudstones, joints which form in the grits will pass upward into the incompetent rocks. The occurrence of well-defined bedding planes between the different rock types may well prevent such propagation, so the author tentatively suggests a second possibility. The sudden relief of distributed strain in the competent rocks by the development of joints will transmit a " shock " to the incompetent material which is capable of forming in the latter substance a system of joints related, but not necessarily parallel, to the system of joints in the competent material. The mechanism envisaged by the author is as follows:—



5a. GENERAL DISTRIBUTED TENSILE STRAIN IN A COMPETENT ROCK.



5b. DISTRIBUTED TENSILE STRAIN RELIEVED BY JOINTING. ARROWS INDICATE DIRECTION AND RELATIVE AMOUNT OF TRANSLATIONAL MOVEMENT WHICH RESULTS.



5c. P, Q, DIRECTION OF GREATEST TRANSLATIONAL MOVEMENT.
AA, BB, JOINTS IN COMPETENT MATERIAL.
XX, YY, JOINTS IN INCOMPETENT MATERIAL.

TEXT-FIG. 5.—Joint development in incompetent material.

The general tensile strain in the competent rock (see Text-fig. 5*a*) increases gradually and suddenly gives rise to joints in the competent material, as indicated in Text-fig. 5*b*. The tensile strains are suddenly relieved and the whole block " contracts " about a central line, represented by *OO*, the periphery of the block naturally undergoing the largest " displacement ". Because of the friction which must exist between adjacent beds, this sudden release of distributed strain in the competent material is transmitted to the shale in the form of a pulse which, if sufficiently large, will initiate tension joints in the incompetent material.

Consider the case when orthogonal sets of " Tension " Joints are formed simultaneously in a competent band. The sets of joints are represented in Text-fig. 5*c*, in plan, by *AA*, *A'A'*, etc., and *BB*, *B'B'*, etc. Each block, at the instant the joints develop, will contract towards the central points *O*, *O'*, *O''*, etc. The amount a point is displaced towards *O* is proportional to the distance of that point from *O*, so that the largest displacement, represented by *PP*, *QQ*, etc., will occur at the corners of each block. Thus, the largest pulses of tensile stress, transmitted to the incompetent material, occur in the *PP* and *QQ* directions which will respectively give rise to *XX* and *YY* joints in the weaker material.

It was assumed that the *AA* and *BB* joints in the competent material were " Tension " Joints which, ideally, will appear in the " movement picture " as " Cross " and " Longitudinal " Joints. It is interesting to note that the development of such fractures in the competent rocks can give rise to " Tension " Joints in the incompetent beds which have an orientation normally associated with " Shear " Joints.

If the distance separating " Cross " Joints in the competent beds is greater than that between " Longitudinal " Joints, then the inferred angle of fracture of the " Pseudo-Shear Joints " in the softer beds will be obtuse, i.e. ($2\alpha \geqslant 90°$). Thus, a commonly observed phenomenon can be explained without recourse to the discredited Strain Ellipsoid Theory of Fracture.

X. Conclusions

From the arguments outlined above it appears that all Master Joints and many Minor Joints, both in competent and incompetent material, are post-compressional phenomena. " Tension " Joints develop in sediments which have not been subjected to considerable tectonic compression, or in which tectonic stresses have relaxed, while both " Shear " and " Tension " Joints may develop in highly compressed and folded rocks. Both types of structures can in some ways be considered as " compression " fractures, but in both instances the actual jointing is brought about by the lateral expansion and

tensile stresses which are attendant upon uplift. The structures are, in fact, not formed until the rocks are within a few thousands of feet of the surface. The relationship of these structures to the " movement picture ", resulting from the original compression, is a consequence of the " residual stresses " which reflect conditions in the tectonic phase. That Shear Joints result from the action of residual stresses accounts for the observed fact that Shear Joints, in general, exhibit no detectable lateral movement. Further, because movement of each joint fracture is so small, a large number of fractures must develop in order to dissipate the residual stresses. The actual joint frequency is determined by the amount of strain energy stored in the rock prior to the onset of jointing.

Hence, the difficulties outlined in the Introduction have been resolved. Joints are the result of both shear and tensile stresses and the mechanics of formation are a natural consequence of their physical properties and the sequence of events to which the rocks are subjected, namely, burial, compression, and subsequent uplift.

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9 ASHTON GARDENS,
HOUNSLOW,
MIDDLESEX.

CORRESPONDENCE

SEPTA AND SUTURES IN JURASSIC AMMONITES

SIR,—It would be futile to prolong the controversy which arose between the late Dr. Arkell and Dr. Westermann concerning the use of septal form in ammonite classification, and in particular the application of this method to the Middle/Upper Jurassic Stephanocerataceae and Perisphinctaceae. The phylogenetic problems will doubtless be solved in time as evidence accumulates. I would like, however, to make some comments on Dr. Westermann's latest rejoinder (*Geol. Mag.*, Nov.–Dec., 1958).

It is helpful of Dr. Westermann to give us in Section B and figs. 1, 2, and 4 of his paper an English version of the septal classification in his earlier paper (1956, pp. 239–243, figs. 1, 2, and Beilage 1), but this does not provide any further confirmation that the septa can be specially relied upon in phylogenetic classification. According to the explanation of fig. 4 the evolutionary sequence of septum-types corresponds to the ontogenetic sequence; in other words, " recapitulation," embraced uncritically by Hyatt and Buckman, *c.* 1870–1930, but found to be unworkable and now generally abandoned by ammonitologists, is now assumed to apply to the septa. It is also stated (p. 451) that evolutionary relationships may be inferred from " the similarities and suggested transformation trends of septal types ", although it is not clear how these trends are discovered before the evolutionary pattern itself has been demonstrated. This kind of procedure, like Buckman's assumed cycles of ornament and other characters, seems to me to be one of the most dangerous kinds of palaeontological method, which can rapidly lead to the erection of a whole structure whose validity depends upon unproved assumptions.

Two other ways of using septa to elucidate phylogeny are mentioned (p. 450–1): (1) " by analogous arrangement on the basis of equivalent septal types." I do not understand what this sentence means, unless it merely means grouping like forms together. (2) " from the similarities of the non-adaptive septal characters." The non-adaptive characters are enumerated (p. 447), but their recognition as such depends on acceptance of the Pfaff hypothesis of septal function, which is not universally accepted (Arkell, 1957, p. 243) and cannot be tested. Arbitrary separation into adaptive and non-adaptive characters is merely liable to lessen or obscure the taxonomic value of the septum or suture-line as a whole.

One particular example may be examined. With regard to the distinction between the subfamilies Otoitinae and Normannitinae, which he places in different families, Dr. Westermann writes that the differences between the suture-lines " correspond with the abullate and aplanulate septal main-types which rank at the family level " (p. 451). The two septal types are those in the " heterochrone " series which are found in sphaerocone and serpenticone shell-forms respectively. According to Pfaff's theory, accepted by Westermann, changes in the septal type must follow automatically on changes in shell-form. One may choose to regard the difference between the shell-forms as the basis of a distinction at family level, but the correlated septal differences cannot add further weight to the distinction. In this case Dr. Arkell believed that the difference between *Otoites* and *Normannites*, which is principally one of shell-form, was of no more than generic importance.

In the case of the origin of the Parkinsonidae (p. 454), does not Dr. Westermann's acceptance of the descent of *Parkinsonia*, with an " aschizolate " septum, from " *Infraparkinsonia* ", with a " euplanulate " one, make nonsense of his own diagram (fig. 4) showing " transformation of septa . . . in ontogeny and evolution "? Reference to this diagram shows that these two types occur in different series of " transformations ".

In conclusion, it seems to me that Dr. Westermann has not answered Dr. Arkell's principal objection to his method: namely that septa can often be demonstrated to be unreliable, and, more important, that there is no

substitute for an evaluation of all characters, taken in conjunction with stratigraphical evidence, in making a classification. He has repeated what he has said in previous works, but he has not advanced any new evidence to convince more conservative palaeontologists that septa, any more than any other single character, are the key to ammonite phylogeny.

D. T. DONOVAN.

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ULTRABASIC PILLOW LAVAS FROM CYPRUS

SIR,—Mr. D. W. Bishopp, who initiated the Cyprus Geological Survey in September, 1950, published the following papers on the Troodos Massif:—

"The Troodos Massif," *Nature*, vol. 169, No. 4299, 22nd March, 1952.

"Some new features of the Geology of Cyprus," XIX Intern. Geol. Cong., Algiers, 1952.

Much of the preliminary reconnaissance work in the Troodos area, as well as some detailed mapping, was undertaken by Mr. Bishopp. I regret therefore that in the brief history of the Cyprus Geological Survey, given in the introduction of my paper, Mr. Bishopp's contribution was not sufficiently acknowledged and also his two papers were not given as references.

The bulk of Mr. Bishopp's letter concerns the Diabase Formation and as my colleagues Mr. L. M. Bear and Mr. R. A. M. Wilson are more familiar with these rocks I prefer to leave any description and discussion to them. Memoirs by Bear and Wilson, based on detailed mapping of parts of the Troodos Massif, are due to be published in the near future; both put forward evidence in favour of an intrusive origin for most of the Diabase Formation.

I. G. GASS.

GEOLOGICAL SURVEY DEPT.,
NICOSIA,
CYPRUS.
15th January, 1959.

THE VISCOSITY OF ROCK-GLASS OF GRANITIC COMPOSITION UNDER VARIOUS PHYSICAL CONDITIONS: A CORRECTION AND AN ADDENDUM

SIR,—In a recent paper on "Granite: some tectonic, petrological and physico-chemical aspects," published in this Magazine (1958, pp. 378–396) I referred on p. 393 to Saucier's determination of the viscosity of rétinite (a variety of pitchstone) under pressure of gaseous water. Saucier found the viscosity to be 10^7 poises at 980° C. with a pressure of gaseous water of 160 bars. Through some error this pressure is wrongly recorded in my paper as 750 bars; the mistake does not alter the conclusions.

Dr. Saucier has kindly drawn my attention to further important investigations of the viscosity of rétinite by his colleague Sabatier (1956). The results of these experiments are of particular interest because they provide a measure of the effects both of increasing pressure of gaseous water and of increasing temperature in lowering the viscosity of natural pitchstone. The following three series of viscosity determinations were made by Sabatier: (1) Nine determinations made at atmospheric pressure under dry conditions showed

that viscosity decreased from nearly 10^{12} poises to slightly less than 10^{10} poises as the temperature increased from 900° C. to 1,060° C.; (2) Five determinations made at 500 bars, with a mean content of about 5 per cent water, showed a decrease from nearly 10^{10} poises to slightly less than 10^{9} poises as the temperature increased from 700° C. to 750° C.; (3) Six determinations at 1,000 bars, with a mean content of about 6 per cent water, showed a decrease from slightly less than 10^{9} poises to slightly less than 10^{7} poises as the temperature increased from 650° C. to 790° C.

Sabatier's experimental results relate to pitchstone which is chemically related to the minimites in Bowen and Tuttle's experimentally investigated synthetic haplo-pitchstones. His results indicate that at 1,000 bars and a temperature between 700° C. and 750° C., conditions under which Bowen and Tuttle's experimental minimite begins to crystallize, the viscosity is of the order of 10^{8} or 10^{7} poises. For melts of minimite composition, with a temperature no greater than 800° C., to have a viscosity as low as that of basaltic melt at 1,400° C., i.e. 10^{2} poises (Kani, 1934), a pressure of gaseous water of the order of 5,000 bars is necessary, i.e. a pressure corresponding to that of the overhead weight of rocks at depths of 10 miles. Only with temperatures of the order of 1,200° C. would the viscosity of melts of minimite composition be reduced to 10^{2} poises or less at a pressure of 1,000 bars (equivalent to the overhead weight of rocks at a depth of about 2 miles).

Even though its viscosity be no more than 10^{2} poises, however, basalt magma does not form net-veins. It thus emerges that Sabatier's viscosity determinations unite with the geological evidence (Reynolds, 1954) in indicating that, at the time of intrusion, the physical state of microgranite (including granophyre) emplaced as net-veins differed from that implied by the normal usage of the term " magma ". If liquid magma were more than saturated with water foam would form and the result would be to increase the viscosity; the fact that foam is highly viscous compared with either of its phases is commonly overlooked. The evidence suggests that the highly mobile antecedent of the microgranite of net-veins was either (1) gaseous, or (2) a disperse system with gas, the continuous phase, acting as transporting agent, i.e. spray. In the latter event, if the temperature were such that the droplets were liquid (saturated with water) a variety of ignimbrite would be expected to remain after escape of the gaseous transporting agent. Ignimbrite does, in fact, occur as net-veins in Ardnamurchan (Reynolds, 1951, 1954, plate 1, p. 590). Furthermore, ignimbrite, like the powdered synthetic glass used in hydrothermal synthesis by Bowen and Tuttle, may be equally expected to devitrify or crystallize under appropriate physical conditions.

Geologists have long been puzzled by the seeming contradiction between (1) the high viscosity of recent acid lavas, which characteristically form tholoids and spines, and (2) the extremely low viscosity implied by the wide areal extent of many pre-historic acid extrusive rocks that were originally described as obsidians and rhyolites. Re-examination of such obsidians and rhyolites of a wide variety of ages and in many countries (including Britain) is increasingly revealing that they are ignimbrites, i.e. they were erupted as " fiery showers ". Reference to one of these recent discoveries of ignimbrite is highly relevant to current discussions concerning the antecedents of granophyre. Over twenty years ago von Eckermann (1936) recorded gradations from felsitic porphyries to granophyres within lava-flows (Dala porphyries) of Precambrian age in central Sweden. Not only does gradation from felsite to granophyre take place through series of progressively older flows, it also occurs within single flows. In examples of the latter spherulitic felsite forms the upper part of a flow and grades downwards into typical granophyre towards the base. In the felsitic portions of some of these flows Hjelmqvist (1956) has recently recognized the diagnostic characteristics of ignimbrite. Even where devitrification has occurred the outlines of the original glass-shards are commonly preserved with the same clarity as in the type ignimbrites from New Zealand (Marshall, 1935). Although unrecognized, and understandably so, at the time when von Eckermann

described these rocks, ignimbrites are well represented in the excellent photographs illustrating his Memoir (von Eckermann, 1936, Plate LXIX, fig. 2; Plate LXXI, figs. 1 and 2; Plate LXXII). Here then is evidence that sheets of granophyre are crystallized ignimbrites; Bowen and Tuttle's experimental results, combined with Sabatier's viscosity determinations, provide evidence concerning the physical conditions under which this could happen without true liquefaction.

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DORIS L. REYNOLDS.

GRANT INSTITUTE OF GEOLOGY,
WEST MAINS ROAD,
EDINBURGH, 9.

22nd January, 1959.

MONOGRAPTUS TRIANGULATUS

SIR,—In my paper, "Triangulate Monograptids from the *Monograptus gregarius* zone (Lower Llandovery) of the Rheidol Gorge (Cardiganshire)," *Phil. Trans. Roy. Soc. Lond.*, B, 241, 485–555, 1958, I established the new species *Monograptus separatus* for a group of forms with rastritiform proximal thecae and triangular distal ones. Seven subspecies were distinguished, depending mainly on the number and nature of the proximal rastritiform thecae present; these were *Monograptus separatus separatus* Sudbury, *M. separatus fimbriatus* (Nicholson), *M. separatus similis* (Elles and Wood), *M. separatus predecipiens* Sudbury, *M. separatus triangulatus* (Harkness), *M. separatus major* (Elles and Wood) and *M. separatus extremus* Sudbury. My work suggested an evolutionary series in which *M. separatus separatus* was the first member and continued as the main stock, and for this reason the group as a whole was given this name.

It has been pointed out to me that according to the Rules of Zoological Nomenclature such a group must always be known by the oldest name available from among those of its members. Thus the species *Monograptus separatus* Sudbury, 1958, should be called *Monograptus triangulatus* (Harkness) 1851, and the type of the group as a whole is the specimen figured by Harkness as *Rastrites triangulatus*, Plate I, fig. 3*a* (? 3*b*) in his paper "Description of the Graptolites found in the Black Shales of Dumfriesshire", *Quart. Journ. Geol. Soc.*, vii, 58–65. The changes of name thus required for the subspecies are listed below:—

M. triangulatus separatus Sudbury, 1958, in place of *M. separatus separatus.*

M. triangulatus fimbriatus (Nicholson, 1868), in place of *M. separatus fimbriatus.*

M. triangulatus similis (Elles and Wood, 1913), in place of *M. separatus similis.*

M. triangulatus predecipiens Sudbury, 1958, in place of *M. separatus predecipiens.*

M. triangulatus triangulatus (Harkness, 1851), in place of *M. separatus triangulatus.*

M. triangulatus major Elles and Wood, 1913, in place of *M. separatus major.*

M. triangulatus extremus Sudbury, 1958, in place of *M. separatus extremus.*

I should like to thank Prof. Dr. Bedřich Bouček for drawing my attention to this matter, and Mr. R. V. Melville for helpful discussion.

MARGARET SUDBURY.

27 AVONDALE ROAD,
MORTLAKE,
LONDON, S.W. 14.

29*th January*, 1959.

INTRUSION-BRECCIA OF DUNMORE, Co. DONEGAL

SIR,—In a recent account of the intrusion-breccia at Dunmore, Co. Donegal, attention is drawn to the association between gas-drilled tuffisitic pipes and post-tectonic plutons (French and Pitcher, 1959). My own work in West Cork has revealed the occurrence of tuff-filled pipes penetrating folded Coomhola series sediments and raises the question of the possibility of the Southern Ireland fold belt concealing a pluton at no great depth.

The West Cork bodies, a complete description of which is in hand, have many points in common with the Dunmore breccia. The most important of these are (*a*) the occurrence of pipes or vents with sharp and steep contacts, (*b*) the emplacement due to gas-stream action, and (*c*) the petrographical similarity of texture and mineralogy of the matrices, except locally where the mineralogy is modified by the mechanical incorporation of comminuted country rock. Certain features of the West Cork bodies are not shared by any of those described intrusions which are more obviously associated with plutons. A significant fact may be that although three plug-like bodies have been recorded the bulk of the intrusions are dykes. Twenty dykes are known with a range in width from a few inches to 15 feet. All these intrusions post-date the main phase of folding. A second distinctive feature of the West Cork plugs is the heterogenity of the fragments and blocks forming the breccia. Graded zones can be defined, (*a*) containing blocks readily identified with the local succession and (*b*) containing exotic blocks and crystals from coarse hornblendite and calc-silicate rocks to gneisses. Single crystals of amphibole up to 6 inches in diameter occur. Thirdly, the degree of metamorphic change is minimal, suggesting that the gas was dry and not hot.

Other evidence suggestive of the proximity of a pluton in West Cork includes :—

(1) The regional metamorphism (which, although never high grade is not simply dynamic).

(2) The sulphide mineralization which elsewhere is associated with granite.

(3) The style of deformation in which cleavage folding predominates. The significance of this is discussed by L. U. de Sitter (1956).

(4) The presence of exposed granite bodies in the easterly continuation of the fold belt in South-West England.

Finally I would like to draw attention to yet another gas emplaced breccia, wholly in granite, occurring at the northern end of the Leinster massif and described by J. C. Brindley (1957).

KENNETH COE.

THE DEPARTMENT OF GEOLOGY,
THE UNIVERSITY,
EXETER.
26th February, 1959.

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REVIEWS

INTRODUCTION TO THE STUDY OF CRETACEOUS OSTRACODA. By H. V. HOWE and L. LAURENCICH. 536 pp., numerous text-figs. Louisiana State University Press, Baton Rouge, 1958. Price $7.50.

This volume attempts to figure and describe every species of ostracod from the Cretaceous System introduced into the literature up to the close of 1956. Some papers which appeared in 1957 are also included, and a two-page appendix annotates a list of supplemental literature. In fact, the authors have achieved a fairly comprehensive survey up to that date, and have even illustrated the twenty general and many more species which in recent years have been described in Russian periodicals by Mandelstam and his colleagues—literature which provincial students in Europe and America still find difficult to consult. The genera are not arranged taxonomically, but alphabetically; the systematic position of each genus is stated if known. This scheme obviates the need for an index, and is easier to use than might be supposed. A 35-page Introduction includes a statement as to the stratigraphical limits of the Cretaceous. Perhaps fortunately the authors have avoided controversy in discussing either the Wealden or Danian in Europe, and have contented themselves by clearly indicating what criteria they have used in determining whether to include species from the marginal horizons. The latter half of the Introduction is devoted to certain aspects of the morphology of the ostracod carapace. Particular attention is paid to the form of the muscle-scars and the nomenclature and classification of the hinge. This section is no mere summary of existing practice, but includes an elaborate nomenclature in which more than half of the sixteen terms used are introduced for the first time. Likewise, the systematic part of the work is not confined to a survey of the literature, and four new genera are proposed. The figures are mainly reproductions of pen-and-ink drawings. They vary much in quality, as is inevitable considering that many are copied from inadequate originals. The majority are clear, and some are quite beautifully executed and a pleasure to study. But few will agree with the authors when they suggest that, in the illustration of the Ostracoda, drawings are very much better than photographs; nor would they necessarily concur with the stipulation that, if photographs are to be used, they should be mounted on a white rather than a black background.

The study of Mesozoic ostracods during the last ten years has gained greatly in impetus, and each year sees the addition of an increasing number of new recruits. The production of the present volume is therefore timely. It will be quite indispensable to all those working with Cretaceous ostracods, and a valuable asset to those with interests confined to the Jurassic or Tertiary. By utilizing the photo-offset process of lithography from a typescript original, costs have been kept relatively low, and the volume is strongly bound in cloth.

P. C. S.-B.

UPPER CRETACEOUS OF THE PACIFIC COAST. By. F. M. ANDERSON, edited by J. WYATT DURHAM. Geological Society of America, Memoir 71, pp. 378 and 75 pls. New York 1958. Price $6.75.

Dr. Anderson devoted a lifetime of study to the Cretaceous of the Pacific coast of the United States and this publication represents his final ideas on the Upper Cretaceous. The first part gives an account of the stratigraphy in California and Oregon and the second part describes and figures the fauna. A tremendous amount of observation is included and the work is an essential source, particularly for the field records.

In view of the authority liable to be attributed to a publication in the Memoir Series of the Geological Society of America, it is a pity that much of the systematic palaeontology is not merely poor, but dangerously misleading. One has to keep in mind that the author died in 1945 and the editor wrote his preface in 1950, both dates well before the publication of the *Traité de Paléontologie* edited by Piveteau or any of the *Treatise on Invertebrate Paleontology* edited by Moore. Consequently the ammonite systematics are bound to be much out of date. Even so there are perpetrated errors such as the use of *Mortoniceras* Meek for the group of *Ammonites texanus* Roemer, a mistake corrected by Adkins in 1928 (Handbook of Texas Cretaceous Fossils. *Univ. Texas Bull.* no. 2838). The editor has allowed the creation of a new genus *Oregoniceras* whilst recognizing in a footnote that this is an invalid junior synonym of *Subprionocyclus* Wright and Matsumoto.

Moreover this work is still another example of the unfortunate tendency to erect new species in order to avoid the troublesome possibility that the author's forms have already been described in another continent. In answer to Nowak's suggestion that *Scaphites condoni* Anderson 1902 might be a synonym of *S. equalis* J. Sowerby, he replies " their identity has not been proved ", instead of showing how they are distinct. Lest it be thought that such criticism only applies to the treatment of the ammonites, one may cite *Echinocorys yolonensis* Anderson n. sp., based on two specimens so poorly preserved that their *generic* determination is admitted to be " tentative ".

Unfortunately the illustrations and descriptions are sometimes too poor to be relied on when considering the possible synonymies: Anderson's material, which is of great importance, will have to be worked over again. The reviewer understands that the ammonites have already been re-examined by Professor Matsumoto.

J. M. H.

PALAEONTOLOGY, Volume I (in 4 parts). Edited by W. H. C. RAMSBOTTOM. Pp. 411, with 72 pls. and numerous text-figs. Palaeontological Association, London, 1957–59. Price 35*s.* per part. Subscription to Association 42*s. per annum* (members receive two parts *per annum*).

A new journal is not always an unmixed blessing. While it may facilitate the exchange of results and ideas within its own specialized field, it often erects new barriers between that field and others, or strengthens those that already exist. This not only increases the scientific isolation of those who regard its field as " theirs ", but also restricts the development of the science by hardening the framework of specialisms to which any new work is obliged to conform.

It is therefore a pleasure to be able to welcome a new journal that is to be devoted not to some new specialism but to a broad field of geological science. The Palaeontological Association, which was founded in London in 1957, has recently completed the publication of the first volume of its journal, *Palaeontology*. In most respects this volume amply fulfils the Association's explicit intention that the journal should cover all aspects of palaeontology. The organisms described in the thirty-nine papers include representatives of every major phylum, and are drawn from almost every geological System. Invertebrates predominate (29 papers), but vertebrates (3 papers) and plant

fossils (7 papers) are also represented. It is pleasing to see no recognition given to that most arbitrary division, between micro- and macro-palaeontology. The great majority of the papers are concerned with systematic descriptions. No doubt this accurately reflects the contemporary interests of palaeontologists, and should rightly continue to bulk largely in the journal; but it is to be hoped that *Palaeontology* will also become a medium of publication for other aspects of the science, such as evolutionary, statistical, and palaeo-ecological studies.

But the most striking feature of the new journal, even at a cursory glance at the volume, is the abundance and fine quality of the plates. Seventy-two plates, interspersed among about 400 pages of text, represent a liberality of illustration which is as admirable as it is rare among palaeontological journals. The plates are collotype, and the reproduction is excellent in almost every case. In a few, however, it is clear that the original photographs were not sufficiently good to make full use of the potentialities of collotype; but no doubt the existence of *Palaeontology* will itself lead, in time, to a greater familiarity with the process and hence to a raising of photographic standards. It is a pity that, on so many plates, excellent photographs are reduced in scientific value by the habit of cutting round the figure before mounting it on the plate; in many cases slightly inaccurate scissor-work by the author and slightly inaccurate blocking-out by the plate-maker have combined to produce outlines that are patently artificial. In fossils in which subtleties of outline are often highly significant (e.g. ammonoids and brachiopods), this practice—followed, one supposes, in the interests of the aesthetic appearance of the plate—is scientifically disastrous. Nevertheless, despite these criticisms, the plates undoubtedly represent one of the strongest features of the new journal, and one that should continue to attract prospective authors. The excellence of the plate-making is well supported by the printing and lay-out of the text, which is attractive and pleasant to read. Text-figures are clear and well reproduced, though in a few of them insufficient reduction has resulted in an unpleasantly coarse texture.

The Editor is to be congratulated on the production of a volume that sets a very high standard for the future; and it is to be hoped that the Association and *Palaeontology* will receive from palaeontologists and other geologists the wide support they deserve.

M. J. S. R.

ELASTIC WAVES IN LAYERED MEDIA. By W. M. EWING, W. S. JARDETZKY, and F. PRESS. Pp. xi + 380, with numerous figs. McGraw-Hill Publishing Co., 1957. Price 75*s*.

Earthquakes, if studied merely as incidents to the normal routine of geological changes, would provide an absorbing study for geologists: indeed, the modern science of seismology owes much to the stimulus given to its development by two British geologists, J. Milne and R. D. Oldham. The special interest of the science, however, arises from the fact that the Earth is an elastic body, and as such transmits elastic waves through its interior and over its surface. The mechanism that engenders the earthquake is associated with the sudden release of a large amount of energy, previously stored up as strain energy in the neighbourhood of the focus of the shock.

The study of the waves through the body of the Earth, has, in the hands of Jeffreys, Bullen, Gutenberg, Lehmann, among others, yielded considerable information about the deep interior. This is a method of great precision, since the velocities of the compressional and distortional waves within the Earth depend only slightly on the periods of the waves, so that a sharp pulse is transmitted to great distances as a sharp pulse. Where the wave velocity depends on the period of the waves the energy of a pulse is received at a distance in the form of a train of waves of varying period: this is the

phenomenon of " dispersion ", and in such circumstances the energy is propagated with the group velocity and not the wave velocity.

The surface waves encountered in the study of earthquakes and artificial explosions are dispersive; their amplitude falls off with depth, and the dependence of wave velocity on period involves the constitution of the part of the Earth's surface across which the waves pass. It might be thought, then, that from a knowledge of the period-velocity relation for a given path one could infer the layering of the region involved. The difference between land paths and oceanic paths does, indeed, show up clearly, but unfortunately the dispersion-curves for a number of different superficial structures are sufficiently alike for their differences to be masked by the lack of precision in the data on surface waves as read from seismograms. Further, there are too many unknown parameters, such as the thicknesses of layers and their elastic properties, to make a unique answer possible.

The method of attacking such problems, therefore, is an indirect one. On the basis of such evidence as is available various assumed constitutions are examined, with the object of comparing observed dispersion curves with those predicted. Thus, a detailed knowledge of the propagation of surface waves in layered structures is one of the basic needs of the study: it is not assumed that in the actual Earth the continents and the ocean floor are uniformly layered, but from a knowledge of the behaviour of surface waves in uniform structures we may expect to be able to examine the more realistic case of a crust whose properties are changing slowly with position on the Earth.

The many possibilities of the propagation of waves along interfaces and free surfaces, where the media may include liquid layers and an atmospheric layer, are comprehensively studied in the work under review. Comparison with the dispersion curves obtained from seismograms shows that the layering of continents and ocean floors agrees, in general terms, with that inferred from studies of near earthquakes and large explosions (both on land and at sea). The balance of the book between theory and observation is admirable, and the many diagrams and illustrations are of considerable help to the reader.

The complete solution of the problem of an impulse applied at a point on or below the free surface of an elastic medium (the so-called " Lamb's Problem ") is known only for certain simplified cases. It describes, of course, the genesis of both body-waves and surface-waves. In many examples in practice the former are less important than the surface-waves, and in this book examples are given of the manner in which the solution is obtained. It is unfortunate that the term " normal modes " has been used in recent years in seismological literature to describe waves associated with a surface, since the expression has a well-understood significance in general dynamics. But there is much to praise and little to criticize in the book, and this lucid exposition of the subject, much of which owes its development to the authors, will be widely welcomed by all workers in this field.

R. S.